TEMPERATURE VARIATIONS OF RUNOFF FROM HIGHLY-GLACIERISED ALPINE BASINS

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By

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Abstract

The temperature of meltwater emerging from the portal of an alpine glacier is usually around 1°C, irrespective of discharge, but warming with distance downstream is influenced by both flow as well as the availability of heat energy. Warmer conditions lead to greater energy availability but also to enhanced icemelt and hence larger quantities of water in proglacial streams. The greater the discharge, the higher the velocity at which meltwater flows and hence the shorter the time available for water to be heated over a fixed distance from the portal. At the diurnal timescale, water temperature rises in the morning before reducing as the amount of water flowing increases in early afternoon. Initially, increasing energy availability is adequate to raise water temperature but fails to keep pace with the rising volume of water in the channel. As a consequence, water temperature declines with continuing increase in discharge. At the seasonal timescale, ice-melt contributions to runoff suppress water temperatures in summer after a spring maximum, giving a distinctive seasonal temperature regime in rivers draining from highly glacierised basins. Highest water temperatures occur at relatively low flows in April and May and then subsequently as icemelt discharge increases with higher air temperatures and rising transient snow line, water temperatures decrease, mean monthly water temperatures being maintained at about the same level between July and September. A paradox appears therefore in that when solar radiation and air temperatures are high, stream water temperatures are often reduced. On warm days, the temperature of the tongue of cool water extending downstream from the glacier portal can be lower and the length of the cool plume longer. Records of water temperature and discharge from rivers draining from two large Alpine valley glaciers in Wallis, Switzerland have been examined. Data for the Massa, draining from Grosser Aletschgletscher throughout the period 2003-2011 have been analysed, and similar hourly measurements were undertaken in summer months on the Gornera, draining from Gornergletscher, both series being obtained close to the glacier termini. As air temperatures continue to warm, this paradox suggests that water temperatures in glacier-fed streams will cool before warming as the deglaciation discharge dividend first increases runoff before subsequent decline. This evolutionary pattern of flow and temperature development can be approached using modelling. Different approaches to modelling are analysed and recommendations made to develop a simple temperature index runoff model, coupled with a deterministic model to predict water temperature.

1. Introduction

Glaciers are regarded as sensitive indicators of climatic variability in high mountain regions (Oerlemans, 1994). Glaciers in high mountain regions across the globe have been in a general state of recession since the Little Ice Age maximum glacier extent (~1850). The process of deglaciation has accelerated since the 1980s as air temperatures have risen and precipitation levels decreased (Barry, 2006). Likely continuous rises in temperature with time impacts on discharge regime and quantity of flow in rivers issuing from highly glacierised basins. Such potential changes are highly important to hydro power installations, irrigation schemes and water resource management. In high mountain regions, such as, the European Alps, 50% of Switzerland's electricity comes from hydro power (Hanggi and Weingartner, 2012; Farinotti *et al.*, 2011).

Warmer air temperatures and potentially high discharges will affect water temperature in mountain streams. Water temperature will be a key indicator of the effects of climatic change on river habitats (Brown *et al.,* 2007). Glacially fed streams are extremely sensitive to climate change. All mountain glaciers in the world are currently being subjected to changing climate, and are experiencing varying levels of melting and generate differing run-off responses (IPCC, 2007). Enhanced levels of melting on glaciers directly affects volumes of discharge entering streams issuing from the ice mass. Increasing air temperatures accompanied by increasing volume of flow directly influences potential positive or negative temperature changes.

Water temperature plays an important role in aquatic ecosystem function (Brown *et al.,* 2007) , there has been growing interest in addressing potential impacts of climate change on water temperature (Caissie, 2006; Fleming, 2005; Hannah *et al.,* 2007; Hari *et al.,* 2006). Water temperature directly affects biotic life in streams through control of energy flow, aquatic organisms metabolic rate and their productivity. Changes in water temperature play a critical role in aquatic ecosystems. The behaviour of cold water fish species has been linked to water temperature (Hari *et al.,* 2006). Any perturbation of the thermal regime of a stream can significantly impact fish habitat (Caissie *et al*, 1998). High mountain catchments are particularly sensitive to water temperature changes because timing, magnitude and duration of runoff contributions from snowmelt, ice-melt, groundwater and rainfall are closely linked to climatic variability (Hannah *et al*, 1999, McGregor *et al*, 1995).

Glaciers moderate flow and decrease year-to-year and month-to-month variation in runoff (Fountain and Tangborn, 1985). In warm, sunny and dry weather at mid-latitudes, increased glacial melt compensates for lack of rain, whereas in cool, rainy or snowy weather, glacial melt decreases (Hock, 2005). This moderating influence was found to be significant in basins with greater than 20% glacier cover. In basins with less than 20% glacier cover, there was no difference in runoff variation from neighbouring basins which receive only snow cover (nival) (Fountain and Tangborn, 1985). Due to storage processes in glaciers, annual peak flow from glacial basins is delayed relative to nearby basins without glaciers (Fountain and Walder, 1998).

Stream temperature is a key indicator of source of flow to a river channel in high mountain areas. In nival basins, radiation has a major impact on stream temperature. This is clearly illustrated when examining water temperature change in regions of streams which are exposed to solar radiation and those which are shaded by riparian cover (Larson and Larson, 1996). It would be logical to believe that the same is true in glaciated Alpine basins. However, it is apparent, from previous studies into high mountain water temperatures (Cadbury *et al.*, 2008; Uehlinger *et al*, 2003; Collins, 2009; Fellman *et al.,* 2013), that as radiation levels and air temperature increase at a seasonal level, stream water temperature remains low with little deviation from the mean in highly glacierised basins. This paradox should be deemed worthy of further analysis. In nival basins stream water is exposed to solar radiation and rises in stream water temperature are observed as radiation levels and air temperature increase (Collins, 2008). Within glaciated basins, however, as solar radiation levels increase so does the rate of melting of ice in the ablation area (Verbunt *et al*, 2003). The melting of ice releases melt water at a temperature close to 0°C, which affects the water temperature, creating an inverse relationship in which, when radiation levels are at their peak stream temperature remains low (Figure 1.1). Hydrological models predict that glacial retreat may result in an initial increase in annual flow. However, prolonged glacier retreat will alter hydrological regime to that of a nival stream, where streams warm and flow decreases as glacier and snowpack sources diminish (Fleming and Clarke, 2005; Stahl and Moore, 2006). Changes in glacier volume that result in declining summer flows could therefore have substantial hydroecological effects, especially for stream temperature (Hood and Berner, 2009).

Volume of flow in a river channel is an important factor affecting variations in stream water temperature. Radiation levels have a significantly lesser impact upon streams carrying a large amount of water (Rothlisberger and Lang, 1987). Larger volumes of melt water lead to a greater discharge, increased velocity, which therefore reduces the time that the water is able to interact with the environment (Brown *et al*, 2006b). The temperature of water entering a river channel is important to the general water temperature downstream, however not as influential as the starting temperature. Water entering the channel at a differing temperature from that which is already flowing within the channel is likely to have some effect upon the temperature. However, this relates to the ratio of the waters entering the channel to those which are already in the channel (Collins, 1979), if only a small quantity of flow enters the channel then it is unlikely to have any effect on stream temperature. An initial shift from positive or neutral glacier mass balance to negative glacier mass balance, melt water generation will increase due to the earlier disappearance of snowpack and the exposure of firn and/or ice, as well as the effects of increased energy inputs (Stahl *et al.*, 2007). However, this initial increase in glacier runoff is unsustainable in the long term because glacier recession decreases glacier area sufficiently to reduce meltwater volumes.

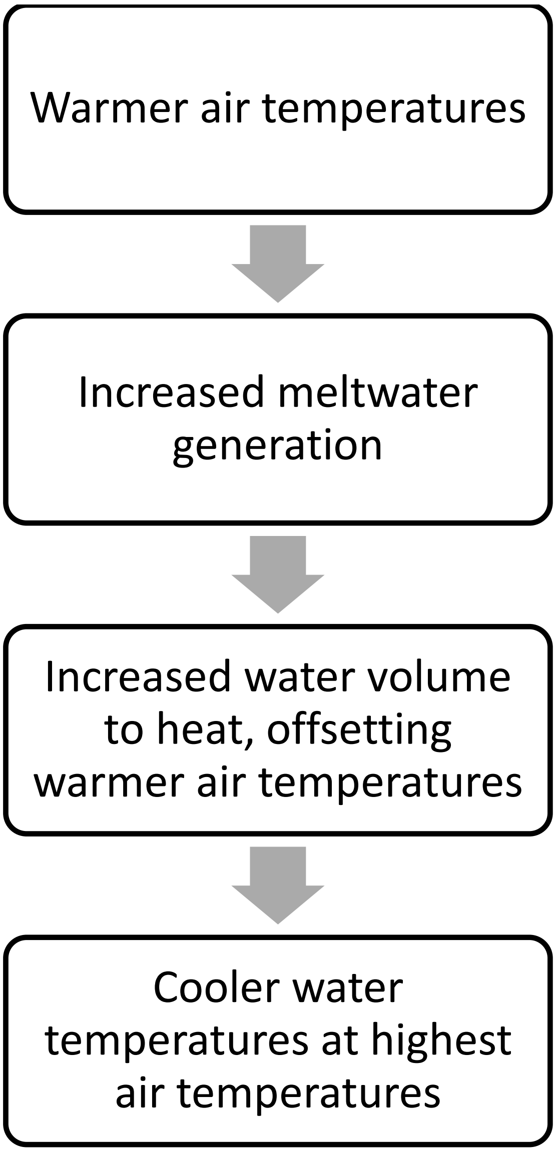

Figure 1.1 Flow chart of paradoxical relationship between air and water temperatures in streams draining large glaciers in high mountain regions

There are annual and diurnal variations of discharge in pro-glacial streams characterised by a lag between air temperatures and discharge (Brown *et al*, 2006a). Winter flow is minimal due to the lack of melting and precipitation falling as snow and not melting. The flow in winter is in the form of groundwater which is held within the rocks and sediments at temperatures above freezing (Ward *et* al, 1999). Spring flow produces large volumes of water, dominated by melting snow pack and exposed glacial ice. Summer flow is dominated by a strong diurnal signal from glacier ice melt, as the transient snowline has progressed to a greater altitude exposing large portions of the glacier. The transient snowline is the lowest altitude of permanent snowpack, and throughout the ablation season the snowline increases with altitude (Jansson *et al*, 2003). The late summer months are often seen to have the greatest flow despite not coinciding with the period of greatest radiation. Ice melt brings large amounts of sediment suspended into the channel. This sediment increases the turbidity of the stream water, reducing transparency (Rothlisberger and Lang, 1987). Turbidity is important as it restricts percolation of sunlight into the depths of the stream and raises stream albedo, hence reducing the rate at which water heats up and maintaining the low water temperature (Richards and Moore, 2011).

Diurnal patterns in highly glacierised basins are characterised by lagged peak events between discharge and air and water temperatures (Figure 1.2). rises in air temperature cause immediate response in water temperatures, as incoming solar radiation reaches the water surface. Increases in discharge are lagged to changes in air temperature, as meltwater generation takes place upstream on the glacier, with flow passing through the basal drainage network before entering the proglacial stream (Rothlisberger and Lang, 1987). Discharge continues increasing to the point where it dominates air temperature as the primary controlling factor of water temperature and as a consequence water temperature peaks before air temperature and discharge. Discharge is further lagged to air temperature, due to the previously explained relationship. However, as the basal drainage network develops over an ablation season, the lag becomes less pronounced as water is delivered more efficiently to into the proglacial environment.

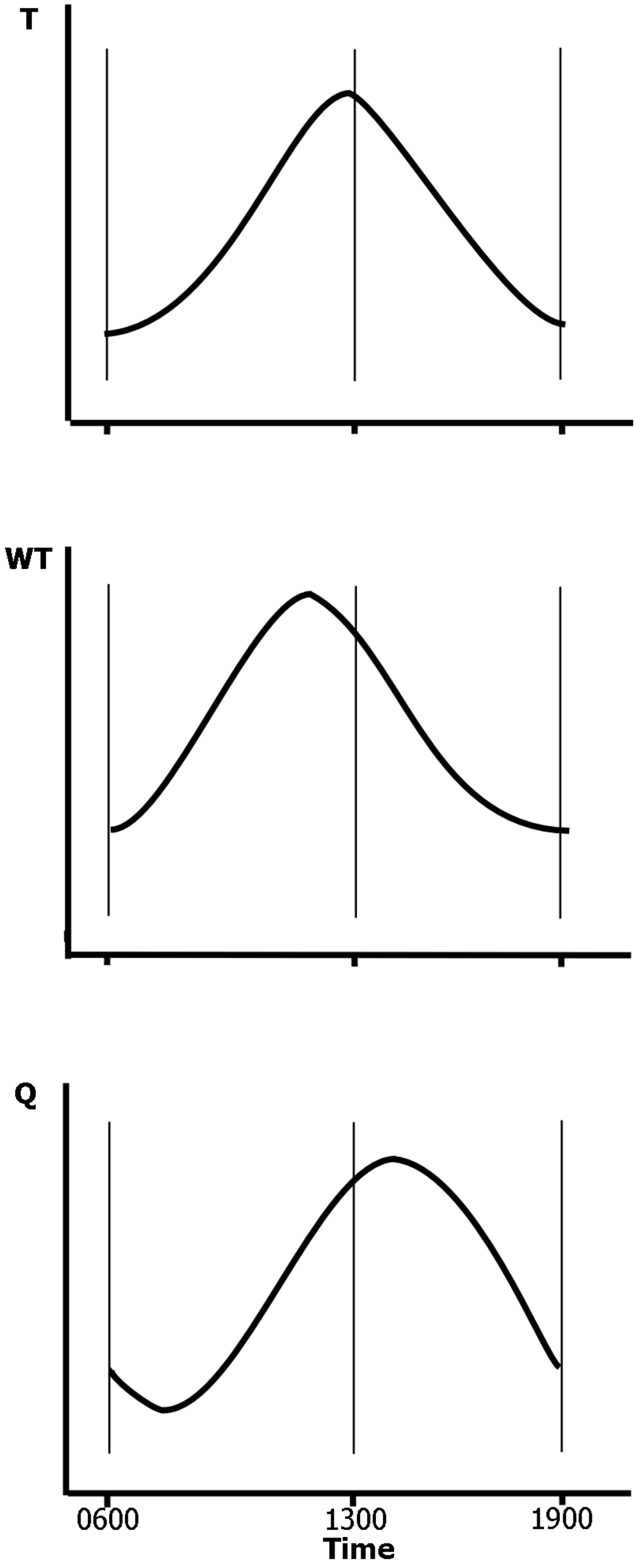

Figure 1.2 Schematic plot of diurnal patterns of air temperature (T), water temperature (WT) and discharge (Q) in a highly glacierised basin during the ablation season

The most rigorous approach to predicting the effects of environmental change on stream temperature is through the use of physically based models that simulate the stream heat budget (Richards and Moore, 2011). Modelling of stream water temperatures has advanced significantly over the past twenty years, with a number of different approaches applied for various situations. However, alpine stream water temperature remains an area where further experimentation is required due to its paradoxical nature with other water courses. In mountain basins there exist undefined spatial boundaries, environmental processes and micro-climates, which all tend to make modelling seasonal discharge complex, inaccurate and error prone (Nash and Sutcliffe, 1970). Modelling can be approached through many different avenues, of type (stochastic, deterministic or statistical) or length (diurnal, weekly, monthly or annually); little research has been undertaken into diurnal modelling within alpine glaciated catchments. Using a glacial temperature index runoff model in conjunction with a proven stream temperature model, simulations of conditions found throughout glacial ablation seasons ought to allow for a much improved understanding of the effects from future climatic changes.

The impact of an altering climate should initially strengthen the paradox between solar radiation levels/ air temperatures and pro-glacial stream temperature. A warming climate ought systematically to increase energy availability for raising the water temperature faster than previous but also increase the rate of run-off generated by glaciers and increase discharge levels. This situation is known to be unsustainable, as it leads to reduction in glacier area and therefore a drop in runoff levels. By assessing the validity of modelling in glacial systems, will allow for future situations under a warming climate and reduced glacier size to be hypothesised. Modelled data can be calibrated using previously recorded data.

1.1 Study area & basin characteristics

Measurements were taken from the gauging station positioned on the Gornera River, the only pro-glacial stream draining from the snout of the Gornergletscher, Switzerland. Gornergletscher is situated at 45° 57 °N latitude, 7 °46 °E longitude with a North-West exposition, in the Pennine Alps, Canton Wallis. It is recognised as the second largest glacier within the Swiss Alps and with a maximum thickness of 450m. Gornergletscher's tributaries are Grenzgletscher, Zweillingsgletscher, Schwarzgletscher, and Breithorngletscher. Theodulgletscher and Monte Rosagletscher have retreated, and no longer contribute mass to the main ice flow but do contribute melt water and sediment.

The basin drained by the Gornera River is 82km² and 55% glaciated, by previously listed glaciers, the elevation range covers 2634m from the peak of Monte Rosa (4634m) to the river gauge, 1.6km from the glacier snout at 2000m giving a stream gradient of 20%. Grenzgletscher sourced on Monte Rosa accumulates cold firn at high elevation (>4000m) all year round (Huss *et al.*, 2007).

The Gornersee is the main source of sudden meltwater release in the Gornera catchment area during the ablation season. The Gornersee, an ice-dammed lake, is situated at the confluence of Grenzgletscher and Gornergletscher (Huss *et al.*, 2007). Created by the retreat of the Monte Rosagletscher, it fills during the early ablation season (May-June), gathering meltwater from the surrounding snowpack and ice melt from the Monte Rosagletscher before draining into the Gornera river later in the ablation season (July- August), often in one large high discharge event or as a sustained period of high discharge over several days until the lake is empty. The effects of the draining of the lake on water temperature are not fully understood as drainage is sporadic and not easily predicted.

Measurements recorded at Massa-Blatten Bei Natters account for the runoff from Grosser Aletschgletscher and its tributaries. Aletschgletscher is situated at 46°47° N latitude, 8°06°E longitude Massa River drains a basin area of 195km², with a glacierisation percentage of 65.9 (Bundesamt Für Umwelt, 2003). The catchment ranges from 4195m (Aletschhorn) to the river gauge at 1446m an elevation range of 2749m, and mean elevation of the catchment is 2945m (Bundesamt für Umwelt, 2003). The glacier snout is at 1800m with the river flowing 2.7km before reaching the gauge meaning stream gradient of 13%.

The two basins (Figure 1.3) were subjected to the same overall patterns of climatic influences, although precipitation in Alpine areas is far from uniformly distributed (Collins, 2009). The ablation season in the Swiss Alps runs from May through to October (Julian Days 120 through 300), with the majority of significant activity occurring in the months of June through to September.

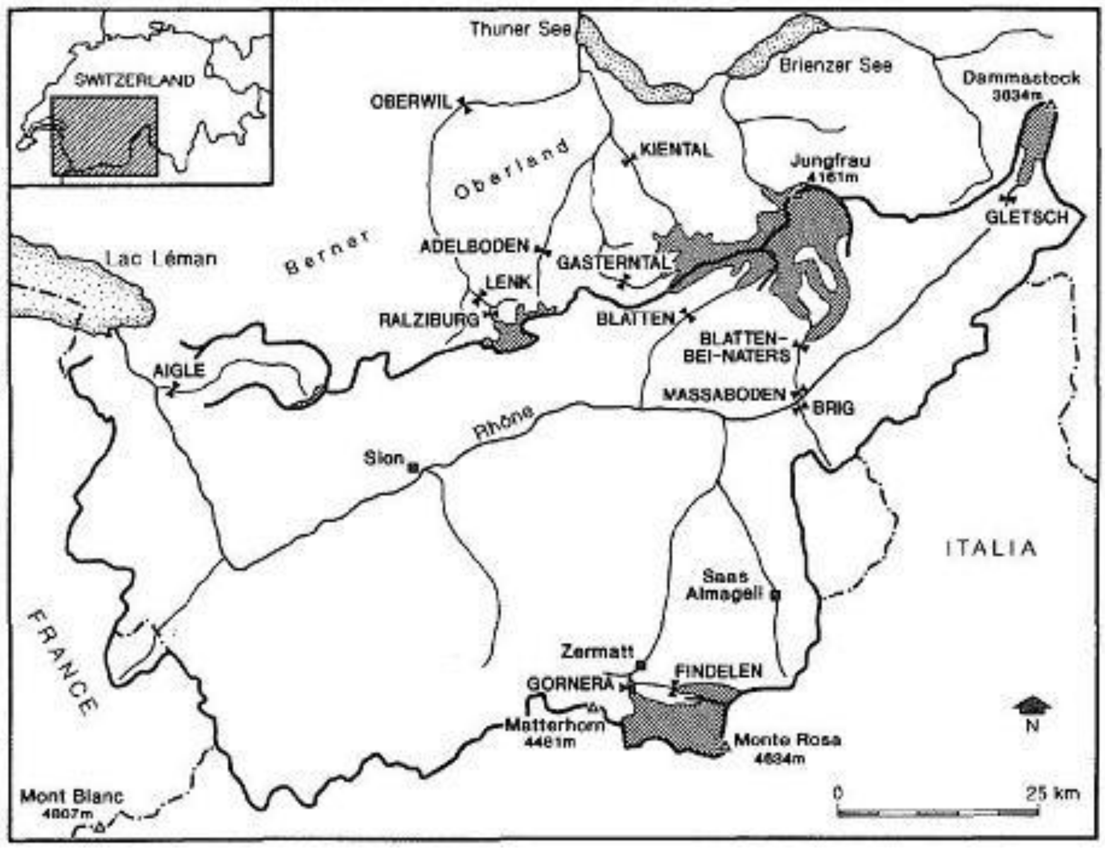

Figure 1.3 Map of Switzerland showing the location of the two basins, Aletschgletscher and Gornergletscher (Collins, Taylor, 1990)

1.2 Aims & objectives

The thesis aims to examine controls on meltwater temperature close to glacial termini in rivers draining highly glacierised basins containing large Alpine glaciers, and to predict how meltwater temperature may respond to climatic change. By exploring the controls on meltwater temperature and examining water temperatures through seasonal and diurnal regimes, primary and secondary influences on meltwater temperature can be identified. Data analysis then allows for the identification of any future trends that may occur during a period of extended glacial recession. Such trends can be extrapolated through the use of modelling, with possible modelling approaches being assessed to understand viability of the thermal regime in a pro-glacial environment close to glacier termini.

2. Background

Understanding the way in which flow is derived in high mountain basins containing large glaciers is integral to hydropower companies, who regulate flow and generate sustainable levels of power all year round. Knowledge of river thermal dynamics is fundamental to understanding instream hydroecological interactions, because water temperature is a key environmental variable influencing physical, chemical and biological processes (Poole and Berman, 2001; Gu and Li, 2002). Physical characteristics influencing stream temperature are spatially nested at: macro- (latitude and altitude), meso- (basin climate and hydrology) and micro- (e.g. micrometeorology, channel geometry, riparian shading and substratum characteristics) scale (Webb, 1996). Specific to high mountain basins with glacial cover is a unique discharge regime, the seasonal evolution of which must be understood in order to understand the impact of discharge on stream water temperature fluctuations.

2.1 Climate change and the Little Ice Age

Ice masses, such as, Polar ice caps and glaciers, are considered to be sensitive indicators of climatic change, responding to both increases and decreases in global air temperatures. Over the past century, global surface temperatures have risen on average by 0.76°C (IPCC, 2007). A slight decrease in air temperature was observed between the 1950s and 1970s. Temperatures have subsequently increased beyond the 1950 levels since the 1980s and predictions indicate that warming is likely to continue into, at least, the near future (IPCC, 2007).

The Little Ice Age is recognised as the period between the thirteenth and mid-nineteenth centuries where air temperatures were lower 'over most if not all the globe' (Grove, 2012). Alongside cooler climatic conditions, some regions possibly received enhanced precipitation levels (Nesje and Dahl, 2003). Colder air temperatures and greater precipitation levels meant mountain glaciers advanced on a large scale. Since the Little Ice Age maximum in 1850, glaciers are estimated to have lost close to 30-40% area and 50% volume (Haerberli and Benitson, 1998). Clear shrinkage and downwasting is evidenced by large terminal and lateral moraines at the Little Ice Age maximum (Collins, 2008).

As global climate has been predicted to continue the warming pattern, this will heavily influence levels of volume change and runoff generation in glaciers (IPCC, 2007). With initial increases in levels of air temperature predicted to generate greater levels of runoff from large high mountain glaciers, as ice melt becomes a greater contributor to annual flow. Initial increases in air temperature will elevate the altitude of the 0°C isotherm, exposing a greater proportion of ice to air temperatures in excess of 0°C, therefore increasing melt generation and runoff. Over time, however, as glacier area reduces so will the amount of melt generated, leading to a decline in total annual runoff (Collins, 2008).

2.2 Discharge patterns

The efficiency of delivery of water into the pro-glacial environment is a great influence upon the ability for water temperature changes, waters with longer transit time have a greater ability to gain heat and solute content than those delivered quickly in bulk (Collins, 1979a). Knowledge of how this subglacial drainage system develops over an annual cycle is important to develop an understanding of how the speed of waters, derived from snow and icemelt, reaching the proglacial system changes over an annual cycle.

The hydrological behaviours of glaciated basins with over 20% ice cover are strongly controlled by the balance between the accumulation and ablation of glacier mass balance (Fountain and Tangborn, 1985). Precipitation in the winter months falls singularly as snow in Alpine environments and temperatures do not reach those required to melt the snow-pack until the spring months, high mountain basins have an extended lag-time. This lag-time exists in all high mountain basins, not only those that are glacierized but also nival basins, nival basins lose the snow cover at the same rate as glacierized basins. However, ice holds much of the meltwater creating an extended lag-time (Collins, 1984). Nival basins therefore have an annual hydrograph that shows large meltwater discharge during the spring months.

Glacierized basins have an extended lag-time, and discharge released throughout the summer months, often held within sub-glacial pockets and ice dammed lakes (Collins, 1984).

Glaciated basins exhibit unique hydrographs in comparison to other nival basins over annual and diurnal scales. These hydrographs differ greatly due to the manner in which proglacial streams gain discharge in comparison with other basins (Fleming, 2005). Hydrographs characterised by glaciated basins in high mountain areas are unique as they are not dependant on groundwater, precipitation or snowmelt as the primary source of flow. Glaciated basins depend on levels of icemelt to generate discharge; atmospheric factors influencing melting therefore control discharge delivery to the proglacial system, peak discharge in glaciated basins occurs in the months of greatest air temperature, where the bulk of glacial melt occurs (Figure 2.1) (Fleming, 2005). Nival basin maximum discharge occurs prior to maximum air temperatures as by this point the seasonal snowpack has reduced significantly. High levels of discharge derived from a cold water source inhibits the potential for large scale water temperature change, creating a cold water pulse during the period where both incoming short-wave radiation and air temperatures are at an annual peak.

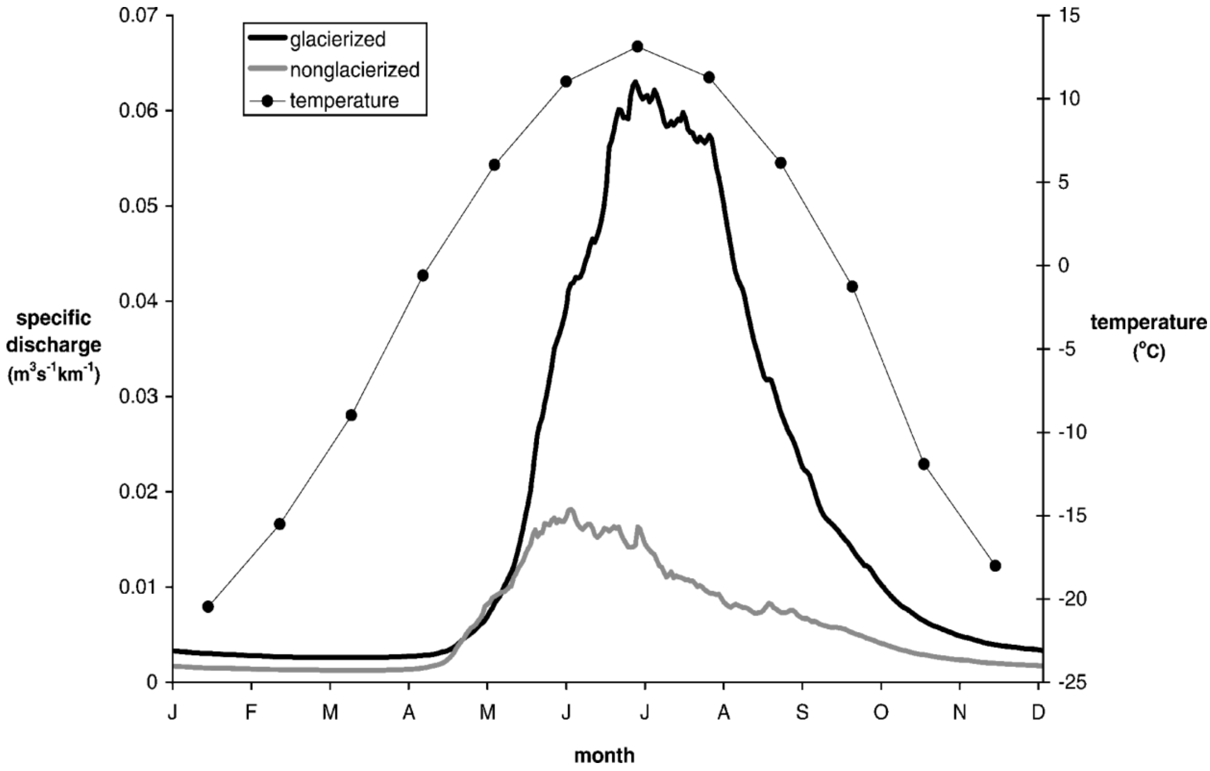

Figure 2.1 Hydrographs of glacierized and nival (nonglacierized) basins (Fleming, 2005)

The annual pattern of discharge in high mountain areas shows that as with non-glaciated Alpine basins, flow in the accumulation season (winter) is composed solely of baseflow (groundwater) and snowmelt dominates through the early part of the ablation season (Collins, 2009). In non-glaciated high mountain basins when snowmelt has surpassed its peak river levels fall and are highly influenced by summer precipitation (Fleming, 2005). In glaciated basins discharge levels continue to rise as runoff is supplemented by melting of exposed glacial ice. Nival basins experience low flow during the advanced latter stages of the ablation season (Aug-Sept) when flow is reliant upon groundwater and precipitation events. During this period glaciated basins experience peak levels of discharge as incoming short-wave radiation, despite being past the annual maximum, and air temperature are high and a large area of glacial ice is exposed to melting (Collins, 1998).

2.3 Development of the subglacial drainage network

During the accumulation season a layer of snow forms on the glacier, this must be melted in the ablation season before solar radiation can reach the glacial ice. The extent to which the snow covers the glacier is represented by the height of the transient snow line (TSL). The TSL will be at its lowest elevation at the end of the accumulation season, i.e. April in the European Alps. Fresh snow has a high albedo of 0.8, while glacial ice has an albedo of 0.4 (Wiscombe and Warren, 1980). Therefore, it takes longer for overlying snow to melt delaying the onset of glacial runoff. The elevation of the glacier and the hydrometeorological conditions, i.e. temperature and precipitation affect the extent to which the TSL reaches, and the time it takes to melt (Figure 2.2). While solar radiation reaches maximum levels on the 21 June in the Northern hemisphere, discharge will not yet be at its peak as the maximum area of glacial ice is not exposed to the atmosphere (Collins, 1998).

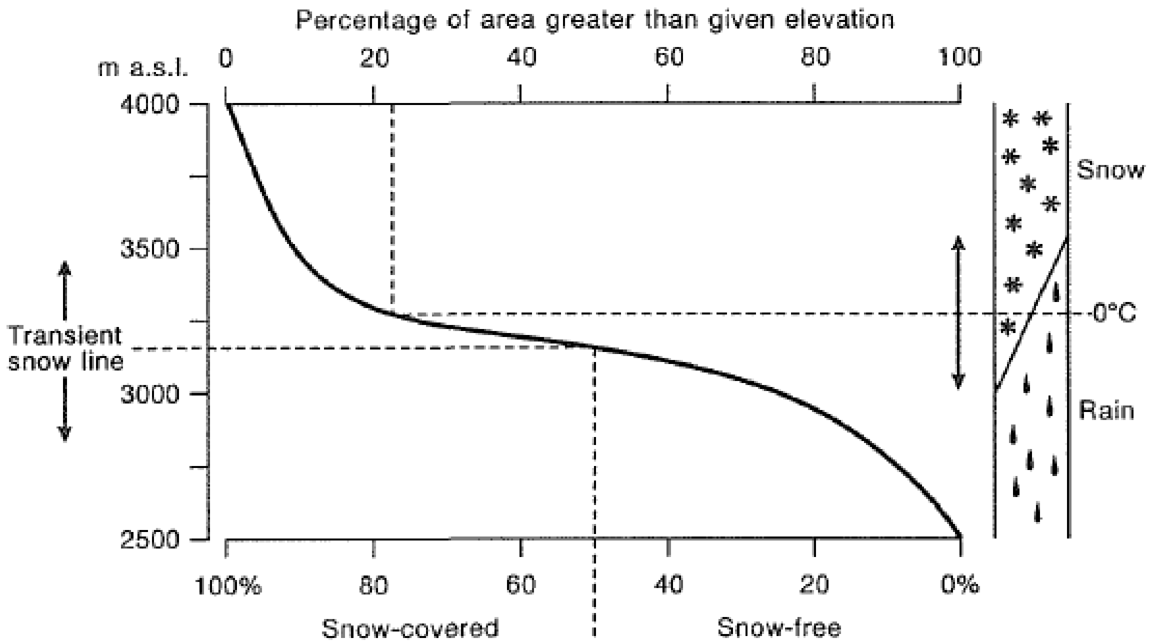

Figure 2.2 Schematic diagram showing how vertical movements of the transient snowline interact with Alpine basin hypsometry to determine the proportions of the basin area which are snow-free and snow-covered, and how the 0°C air temperature isotherm similarly interacts to partition basin area into portions over which precipitation falls as snow and rain (taken from Collins, 1998)

The altitude to which TSL reaches depends on the hydro-meteorological conditions that occur over an accumulation season, this influences the time it takes for annual discharge to peak. The meltwater draining Gornergletscher, Switzerland generally reaches maximum during August, as there is a balance between high incoming solar radiation levels, the altitude of the TSL and the distance meltwaters travel through the basal hydrological network (Collins *et al.* 1989; Swift *et al.* 2005). TSL is at its highest altitude at the end of an ablation season (approx 21 September), but incoming solar radiation levels are much lower, restricting levels of glacial melt generation.

Discharge sourced from a glaciers surface primarily flows through the subglacial network beneath a glacier called the basal hydrological network (Collins 1979c; 1989; 1991; Boulton *et al.* 2001). The basal hydrological network, or subglacial drainage network, develops and changes throughout the ablation season. At the start of the ablation season the basal drainage system consists of a diffuse network of many conduits that cover a large proportion of the glacier bed, these are called distributed channels, and exists when surface meltwater inputs are low, therefore flow velocity is low (Swift *et al.,* 2005; Collins 1989; Jobard and Dzikowski 2006). As the ablation season progresses surface meltwater inputs increase, accessing the glacier bed via moulins and crevasses. Water pressure is raised and conduits expand, and may collapse into one and other to form larger conduits at the expense of the smaller conduits (Hallet *et al.,* 1996). This leaves a channelized system to develop in place of the distributed system (Warburton, 1990). The hydraulic efficiency of the basal drainage network is important as it controls the capacity and direction of flow (Swift *et al.,* 2005).

There are three methods of drainage within a glacial system; supraglacial (above the ice), englacial (within the ice) and subglacial (beneath the ice). The majority of flow is in the form of englacial and subglacial channels, created by the erosive power of flowing water (Hubbard and Nienow, 1997). However, little of the water carried englacially is formed in these tunnels; the majority is supraglacial which has percolated through the ice in moulins and crevasses (Rothlisberger and Lang, 1987). Water carried both englacially and subglacially erodes and melts ice due to an enhanced temperature of ~0.2 °C; this temperature is enough to melt tunnel walls, expanding channels therefore enhancing quantities of water travelling through the system (Isenko *et al,* 2005). It is recognised that water entering the glacier from a static body of water open to climatic radiation (ice-dammed lakes) would be capable of increased erosion before being brought to the equilibrium temperature (Isenko *et al,* 2005). Supraglacial channels are influenced by gravity and therefore, tend to flow into the englacial and subglacial channel systems once they have sufficiently eroded the surface ice (Figure 2.3). Subglacial lakes form in areas where the hydraulic potential is relatively low, and englacial pockets generally form as remnants of crevasses or drainage tunnels (Rothlisberger and Lang, 1987). It is these features, which restrict the flow of meltwater through a glacier and allow for major build-up and sudden release.

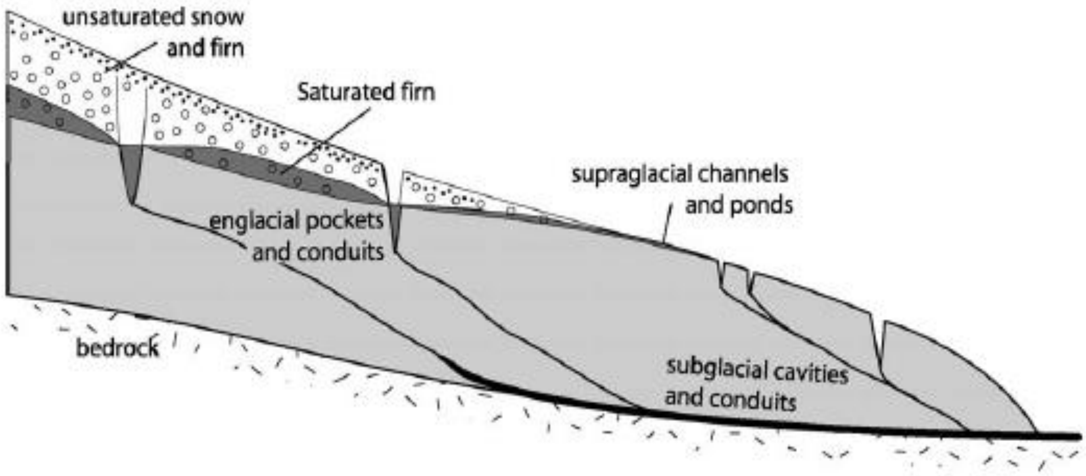

Figure 2.3 Methods of water transport within a glacier (Jansson *et al.,* 2003. Adapted from Rothlisberger and Lang, 1987)

During passage from a glacier meltwater can follow various flow pathways. Meltwater can flow into the pro-glacial stream through runoff or englacial passageways within the glacier (Collins 1991). Moulins and crevasses act as a tunnel for glacial melt to pass from the surface to the glacier bed and flow through the cavities of the subglacial hydraulic system into the stream (Gurnell *et al.* 1996; Clifford *et al.* 1995). Huss et al (2007) carried out a study on Gornergletscher, Switzerland and found that, unlike Jansson *et al.* (2003), glacial conduits are more likely to reach to the rock surface under the glacier, allowing meltwater to flow through these passageways and be transported underneath the glacier to the proglacial stream. Glaciers generate meltwater by processes of basal melting and surface ablation (Swift *et al.* 2002). The flow pathway of meltwater has a critical influence on the development of the subglacial drainage system. If the concentration of surface meltwater is high, and adequate capacities reach the glacier bed it is more likely that a channelized system will form with few large arterial conduits (Brown *et al.,* 1996).

2.4 Influences on Biota

Despite the importance of thermal conditions in influencing biodiversity of high mountain river systems, knowledge of year round stream temperature variability is limited (Brown *et al.*, 2006). These limits become increasingly clear when attempting to attribute climatic factors to temperature change in high mountain streams draining from the terminus of glaciers. Climate change will directly affect runoff levels in high mountain streams draining from glaciers. River thermal patterns will be modified by any changes to discharge, and so the effects of water temperature change on biotic habitats are of high importance. Knowledge of river thermal dynamics is fundamental to understanding instream hydroecological interactions, because water temperature is a key environmental variable influencing physical, chemical and biological processes (Poole and Berman, 2001; Gu and Li, 2002).

Biotic life in high mountain streams close to glaciers primarily takes the form of micro-organisms. Bacteria, viruses, fungi and protists are known to flourish on bare rock surfaces (Rott *et al*., 2006). Combined with algae these micro-organisms are referred to as benthic biofilm (Milner *et al*., 2009). Aquatic life forms have a specific range of temperatures in which they can survive; Logue *et al*. (2004) observed lower biomass of bacteria in meltwater issuing from glacial basins than those emerging from groundwater-fed streams in the Swiss Alps. Changes in temperature brackets for existing aquatic life forms in proglacial streams can result in habitat restriction or destruction (Caissie *et al.*, 2007). Changes in the diversity and abundance of micro-organisms under scenarios of altered cryosphere hydrology could potentially have widespread implications for glacial river ecosystem food webs (Milner *et al.*, 2009). Expected long-term increases in stream water temperature with glacial retreat can be expected to drive quicker organic matter decomposition and nutrient cycling. Changes in the biogeochemistry of alpine streams with a shrinking cryosphere will likely encourage heterotrophic action (Milner *et al.*, 2009).

2.5 Stream temperature influences

Stream temperature varies diurnally and seasonally in response to natural cycles of solar radiation, air temperature and discharge. Stream temperature in a system is influenced at a number of different scales, macro-scale (altitude and latitude), meso-scale (basin climate and hydrology), and micro-scale (channel geometry, shading and substratum characteristics) (Webb, 1996). Water temperature drivers (Figure 2.4) are external to the system and assist in forming a streams physical setting. These factors are capable of controlling rates of heat and water delivery to the system, and are therefore capable of influencing water temperature in a positive or negative manner (Poole and Berman, 2001).

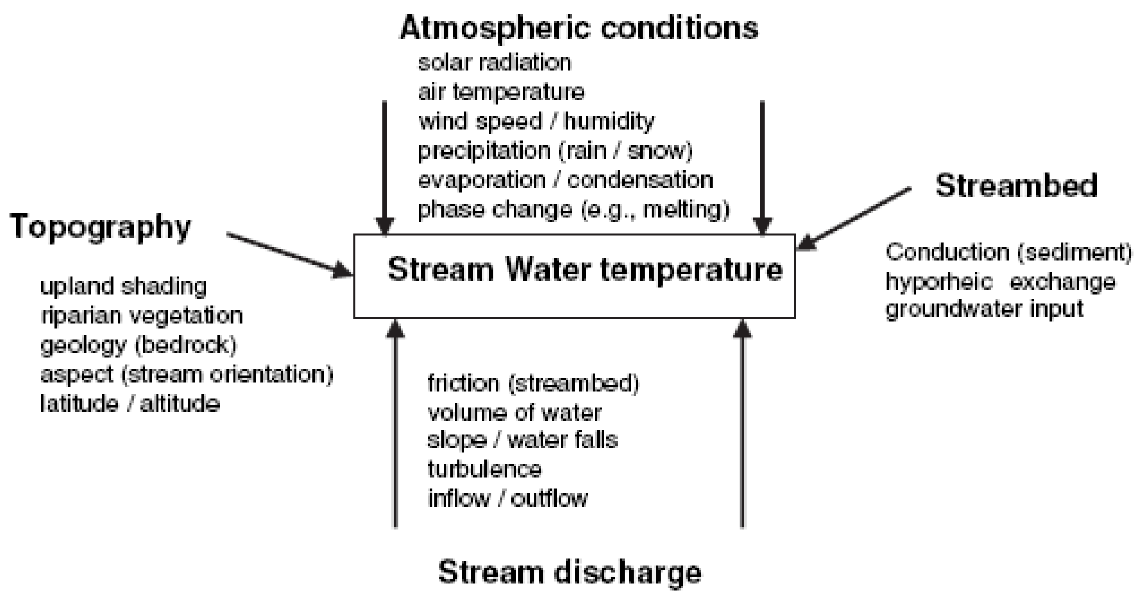

Figure 2.4 Factors affecting stream water temperature (Caissie, 2006)

2.5.1 Water sources

Alpine streams gain discharge from a variety of different sources each with its own distinct thermal signature (Cadbury *et al*, 2008). Water in proglacial streams is sourced from icemelt, snowmelt, groundwater flow and/or precipitation. Each source is not only thermally unique but chemically different making it more easily identifiable by using specific conductivity (electrical conductivity) in conjunction with water temperature (Table 2.1). Ice melt and groundwater are rich in solute content due to their greater residence time and extended contact with bedrock and unconsolidated sediments (Hodson and Ferguson, 1999), allowing them to become chemically enriched. Snowmelt and precipitation are derived directly from the atmosphere and therefore are delivered to the proglacial system having little contact with bedrock or chemically rich sediments before entering the main river channel (Hodson *et al.*, 1998).

Table 2.1 Chemical and temperature characteristics of high mountain stream sources

Water Source	Temperature characteristic	Chemical characteristics
Icemelt	Low temperature	Solute rich
Snowmelt	Low temperature	Solute poor
Precipitation	High temperature	Solute poor
Groundwater	High temperature	Solute rich

The sources of stream water flow and their respective ratios control initial starting temperature in the stream headwaters and control the scale of temperature change downstream. Waters sourced glacially rely primarily upon ice and snowmelt for the bulk of discharge, therefore the greater the percentage basin of glaciation, the greater the control of stream temperature regime by the icemelt and snowmelt (Chikita *et al.*, 2010). Accumulation season discharge consists solely of groundwater as precipitation falls as snow and air temperature remains too low to generate any melting of the snowpack. During the early part of the ablation season, flow consists primarily of melt from the snowpack, before exposing glacial ice. As solar radiation levels and air temperature increase, icemelt supersedes snowmelt to become the dominant driver of discharge until the end of the ablation season where by groundwater becomes the sole supplier of discharge as the accumulation season begins (Rothlisberger and Lang, 1987). During the ablation season, the elevation of the 0°C isotherm indicates over which proportion of the basin rain and snow will fall during precipitation events (Collins, 1998).

2.5.2 Percentage glacierization of basins

Runoff variability in glacial basins increases as glacierization percentage decreases (Collins, 2006). Any changes to runoff regime impact on potential changes in water temperature in proglacial streams. Collins (2008) compared four glacial basins of various levels of glacial cover and a nival basin over a period of 80 years, finding run-off in (near-) ice-free basins reflected fluctuations in precipitation and the greater the glacier cover the greater the influence of thermal conditions. Glaciated basins receive their bulk discharge from ice-melt over an annual period, whereas annual nival basin runoff comes from the melting of winter precipitation and summer precipitation events.

Previous studies into water temperature of glacierized basins (Cadbury *et al.,* 2008 Uehlinger *et al.*, 2003 Brown *et al.*, 2006a; 2006b) where the control of the percentage of glaciation on stream temperature has been investigated have focussed on basins which have ≤30% glaciation, and the basins examined in this thesis are above 50% glacierized. Collins (2009) states that greater percentage glacierization of a basin, the greater the level of control over the annual runoff regime, as the bulk of discharge is composed of snow and icemelt sourced on the glacier.

2.5.3 In-stream influences

In-stream influences on water temperature are directly affected by the discharge level. As the volume of water in the channel increases so does the velocity of the body of water therefore reducing the time between glacier portal and gauging station, in which heating processes can take place.

The effects of stream gradient on water temperature have been demonstrated, by Richards and Moore (2011), to be a contributory factor in raising stream albedo levels alongside changes in discharge. The albedo of a stream can be altered by the greater frequency of "whitetops" and white water in fast flowing steep streams (gradient of>10%) where the bed is characterised by large boulders. Prior to the study by Richards and Moore (2011), whitetops and their relationship to albedo levels had only been studied in oceanic/maritime conditions. Albedo is the reflectivity of a surface, measured between 0 and 1, with 1 being total reflectivity and 0 being total absorption (Wiscombe and Warren, 1980). The albedo of water dictates the level of radiation energy absorption and therefore the potential for water temperature increase.

Shallow gradients of proglacial streams often lead to the braiding of streams or the formation of proglacial lakes, both of which create a much greater capacity for heating than single-channel and steep proglacial streams (Uehlinger *et al.*, 2003). Braided channels are wide and shallow, and discharge is low in individual braids. However, the whole constitutes the full level of ice melt. Shallow, wide channels absorb energy quickly as velocity is lower and surface area increases, compared to single channel systems, allowing for greater percolation of solar radiation.

Turbidity is a measure of the haziness of a fluid caused by individual particles; the level of turbidity determines the capability of light and solar radiation to percolate the water surface. Glacial rivers have high-suspended sediment loads (Collins, 1979, Hodson *et al.*, 1998) during periods of high discharge during the ablation season. High sediment concentration intercepts incoming short-wave radiation, reflecting and refracting radiation (Richards and Moore, 2011). The greater the level of percolation, the greater the potential for temperature change. Particles carried in suspension in streams reflect solar radiation back into the atmosphere, increasing the albedo of flow. Proglacial streams are heavily sediment laden when carrying glacial meltwater, due to the high levels of erosion by glaciers at the ice-bedrock interface (Collins, 1979b).

2.5.4 Atmospheric influences

Atmospheric influences of stream temperature are the key drivers for temperature changes as they are responsible for heat exchange processes at the water surface (Figure 2.5). The main atmospheric factor driving temperature change is solar radiation (incoming short-wave radiation); other atmospheric factors influence solar radiation, indirectly affecting stream temperature such as, cloud cover.

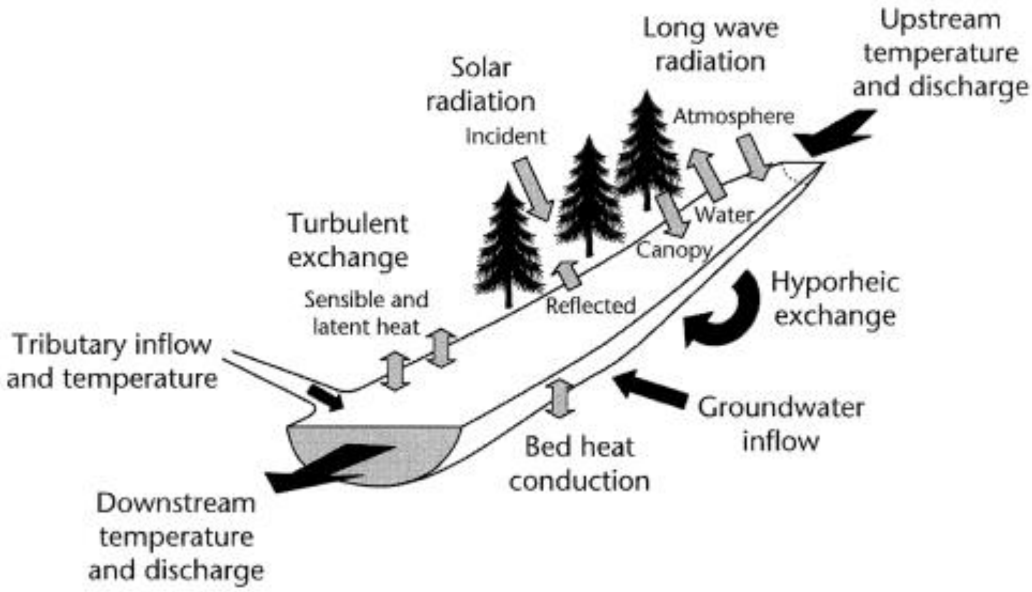

Figure 2.5 Factors controlling stream temperature. Energy fluxes associated with water exchanges are shown as black arrows (Moore *et al.*, 2005)

Energy availability determines the levels of runoff generated through ice and snowmelt and the level of temperature change in discharge. Annual maximum solar radiation in the Northern Hemisphere occurs on June 21 and the minimum on December 21, peak air temperatures are lagged from this maximum as earth's atmosphere takes longer to heat therefore maximum temperatures frequently occur between late July and August. Heating at the water surface occurs through direct incoming short-wave radiation and through convection and conduction from the atmosphere (air temperature) (Figure 2.5) (Moore *et al.,* 2005). Influences which affect direct incoming short-wave radiation are explored further in this chapter.

Precipitation influences stream temperature as it inputs large quantities of water into the system at a different temperature to that which is already flowing in the stream. Precipitation falling in high mountain Alpine areas will have a temperature greater than that of water flowing in the channel and be closer to that of current air temperature, however there is a lack of understanding of the impacts in highly glacierized Alpine areas and have only principally been covered by Collins (1998). For precipitation to occur at any time there must be adequate levels of cloud coverage. This cloud coverage subsequently affects direct solar impact radiation levels, the primary factor driving melting of ice and snow (Jansson *et al.,* 2003). In the accumulation season, precipitation falls as snow, and remains this way until the 0° isotherm advances in elevation above the lowest reach of the catchment which will receive rainfall (Collins, 1998). Storm systems in mountainous regions originate quickly and often due to the relief of terrain. These storms form as rising hot air meets descending cool air. Due to the nature of summer storm events in Alpine areas, large quantities of water enter the river system in a short space of time creating a pulse of warmer water in a cool water system.

The timing of storms in Alpine areas has a greater effect on stream temperature when occurring at different diurnal and annual timeframes (Collins, 1998). Diurnally impacts of precipitation will have the greatest effect on stream temperature during periods of low flow from other water sources, where therefore the precipitation will account for a greater percentage of river flow and have a greater impact in changing water temperature. Precipitation events during peak diurnal flows are likely to have a much-reduced effect upon stream temperature as the event accounts for a much lower proportion of total flow.

Impacts of storms at an annual timescale are more a consequence of the state of development of the glacier basal drainage network, which controls the delivery of ice/snowmelt to the proglacial stream system (Collins, 1998). Early in the ablation season when the drainage network is in its infancy, undeveloped and consisting of linked cavities, the lag-time for the water to reach the proglacial stream is long (Jansson *et al.*, 2003). Precipitation falling on the glacierized area of a basin will undoubtedly contact the ice and will therefore have its temperature altered toward 0°C; the length of time water is in contact with the ice will impact upon the amount of temperature change. Precipitation high on the glacier, a long distance from the snout and portal will have a lower temperature than precipitation falling close to the snout; the temperature of precipitation is controlled by the elevation of the 0°C isotherm in the atmosphere. Position of the 0°C isotherm on the glacier determines the state of precipitation (snow or rain), rain falling close to the 0°C isotherm on the glacier will have been held in contact with the ice for a longer period and have been altered closer to freezing point. Later in the ablation cycle, lag-time is reduced, as the basal drainage network will have developed from linked cavities to conduits becoming more efficient delivering water with less delay to the proglacial system. Ice contact time is diminished and water temperatures remain further from freezing point.

The overall impact of precipitation events as stated by Collins (1998) depends on the proportions of glacier area free of snow, the positions of the 0°C isotherm and transient snowline and the development stage of the subglacial drainage network.

2.5.5 Shading

Solar radiation is very much a function of site conditions and predominately related to the degree of shading (Hebert *et al.*, 2011). Shading, whether riparian, topographical or cloud cover, is the greatest determinant for stream water temperature change as it directly influences the levels of solar radiation available at the water surface.

Riparian shading of water courses is considered a major factor in intercepting direct solar radiation from reaching the water surface (Johnson, 2004). In high mountain regions vegetation has neither the density nor height necessary to shelter streams from incoming solar radiation. Riparian growth is restricted due to shallow and particularly, in glacial valleys, nutrient poor Alpine soils that lack organic compounds. Alpine Tundra climate restricts the length of the growing season due to a combination of low winter and night temperatures and, permanent snow cover during the accumulation and early ablation season (Johnson, 2004).

Topography influences stream temperature by shading a watercourse from direct solar radiation reaching the water surface, topographical shading not only accounts for large landforms but also river banks (Larson and Larson, 1996). Shading is influenced by the orientation of a stream and also the latitudinal position. The path of the sun changes throughout its annual cycle from 23.5°N on 21 June to 23.5°S on 21 December. The influence of topography is greater at times when the sun is at its greatest latitudinal distance from the stream, as the angle of reflection is at its most acute from the solar maximum. Topographic shading is most prevalent in narrow streams with high bank sides and/or steep surrounding topography (Larson and Larson, 1996). In glacial basins, shading is not confined solely to the proglacial stream but also to the glacier itself. The glacier can be thought of in similar terms to a river channel with the flanks of the glacier receiving the greatest level of shading from steep valley sides.

2.6 Summary

In summary, it is clear that a number of factors, some common to all rivers in temperate regions and some, which are only a focus in high mountain environments, directly influence stream water temperatures in rivers, which drain from highly glacierized basins. As water temperatures of flow entering the pro-glacial environment are close to 0°C, it is likely that all the factors detailed will impact on the degree of temperature change, with primary focus being placed on incoming short-wave radiation, air temperature and discharge levels.

3. Methods

3.1 Data collection

3.1.1 Meteorological measurements

Climatic measurements were taken from a number of locations around the Gornergletscher and Grosser Aletschgletscher systems. Gornergrat meteorological station, which provides data of all meteorological parameters around Gornergletscher, is situated at 7°47'9"E, 45°59'4"N, at an altitude of 3110 m.a.s.l. Sion Aeroport meteorological station provides data for the Massa basin (Grosser Aletschgletscher) and is located at 46°21'94"N, 7°32'67"E. Air temperatures were measured 2-metres above ground in a sheltered environment. The method prevents any backscattering of surface heat and protects a sensor from direct incoming solar radiation. Air temperatures vary with altitude and for the purposes of this thesis were corrected for elevation differences between the meteorological stations, gauging stations and average basin altitude (-1110m for the Gornera and +964m for the Massa) using the adiabatic lapse rate (Johnson, 1999):

$$\mathbf{Ta = To +/- Ct * (Z\text{-}Zo)} \qquad \mathbf{(3.1)}$$

where **Ta** is representative of air temperature at the mean elevation of the segment (**Z**). **To** indicates the air temperature recorded at the weather station (**Zo**) and **Ct** represents the moist-air adiabatic lapse rate (-0.00656°C/m). Air temperatures collected by meteorological stations are from 2m above ground to minimise effects from reflected heat and heat emitted by the underlying topography, sensors are shielded from direct solar radiation and wind. Air temperatures at the gauging stations are higher than those that are higher up the catchment and therefore an average elevation for each reach was calculated to best represent the mean conditions across the length of the segment.

In order to analyse data across such a large time scale over numerous annual alpine cycles, air temperature was used as a surrogate to incoming short wave radiation as data were readily available for the full time period and incoming short wave radiation data were unavailable close to the studied basins. The author recognises that there are differences between levels of incoming short-wave radiation and air temperature, not at least that when radiation levels are zero during the night air temperatures can remain above 0°C. However, Chikita *et al* (2010) showed that although short-wave radiation correlated more strongly with water temperatures than air temperatures, the two variables showed the same general trend (Figure 3.1). A lag exists between the annual peaks of incoming short-wave radiation and air temperature, the peak of radiation occurring on 21 June in the Northern hemisphere. Air temperature follows sometime after as the earth's surface absorbs heat over long periods before releasing it, later in summer, as a latent heat store creating a buffer of air temperatures (Chikita *et al.,* 2010).

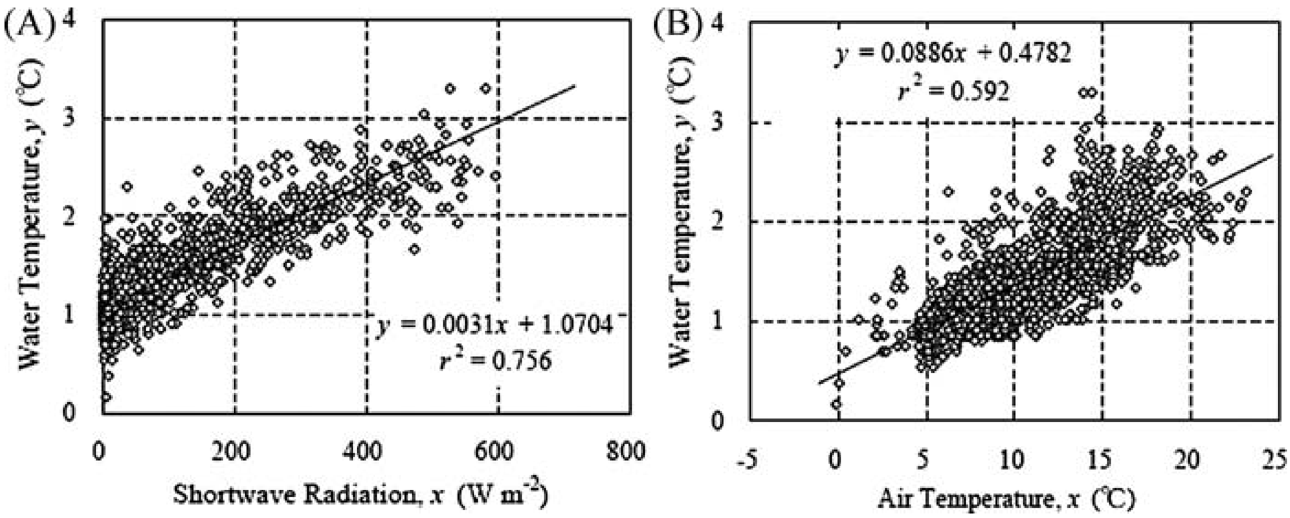

Figure 3.1 Relationships between shortwave radiation (A) and air temperature (B) and water temperatures (Chikita *et al*, 2010)

3.1.2 Discharge measurements

Discharge measurements on the Gornera River are collected automatically by Grande Dixence, a hydroelectric power company and Massa data were collected by Bundesamt für Umwelt. Data were collected at hourly intervals on the half hour (Gornera) and on the hour (Massa) every day. The Gornera and Massa gauging stations channel river flow into a managed single channel, where wetted perimeter and velocity of water are measured and discharge calculated. As water measurements on the Gornera and Massa Rivers are primarily influenced by ice and snow melt, during the accumulation season (December through March) river flow is close to zero and made up solely of groundwater flow, so the focus of study at a diurnal and seasonal scale is on the ablation season (May-October).

Limitations

The gauging station on the Gornera River in periods of high discharge required to be degravelled, in order to remove boulders and stop gravel beds forming at the base of the managed gauge. This process involved the opening of a gate at the base of the gauge allowing water to drain at a greater velocity removing the build-up of gravel. Degravelling occurs twice daily at 0530 and 2030 causing discharge levels to drop and give false readings, for this study, these data have been removed.

3.1.3 Water temperature measurements

Water temperatures were collected under two differing regimes at the Gornera and the Massa. Water temperature was measured at the Gornera gauging station using a MiniSonde 4a probe, the capabilities of which are shown in Table 3.1. Data were recorded at high resolution (10-15min periods) throughout the ablation seasons of calendar years 2004 through 2011. As data were recorded at such high resolution, data sets are limited to short time periods and do not cover whole ablation seasons.

Table 3.1 Properties of MiniSonde 4a probe (Hydrolab, 2013)

	Range
Temperature	-5 to 50°C
Specific Conductivity	0 to 100 µS/cm
pH	0 to 14
	Performance
Memory	120000 measurements
Battery supply	8AA batteries
Battery life	114 days
Maximum depth	225 m

Water temperature recorded at the Massa with hourly resolution throughout the calendar years of 2003 through 2009 by Bundesamt Für Umwelt (BAFU). Data collected at the Massa gauging station are available throughout the year allowing for more in depth annual trends to be identified over the Gornera basin.

Limitations

Water temperature measurements collected at Gornera gauging station were manually collected, using a MiniSonde probe. Probes were installed on field visits in the ablation season, each time during times of high specific discharge (Figure 3.2). Due to high variability of discharge in glacial basins and the wide range between high and low discharges in the ablation season meant that at periods of sustained low discharge the probe became exposed to atmospheric factors and recorded air temperatures rather than water temperature.

Experimental design at the Gornera aimed to primarily at collecting high-resolution electrical conductivity data, with water temperature, pH and dissolved oxygen collected merely as a consequence.

Figure 3.2 Gornera gauging station at high discharge

3.2 Overview of study period

Continuous hourly records of discharge from Massa were processed for calendar years 2003-2011 and Gornera 2004-2008. Water temperature data were obtained for the full measured period for Massa. However, data for Gornera were only continuous over the lifetime of the MiniSonde probes and collected for monthly periods coinciding with scheduled field and site visits during the ablation season. Air temperatures were available close to both gauging stations, for the same periods where discharges were collected. Annual patterns of data were examined through 2005 as this was an accurate average of the period studied, diurnal regimes were analysed across a variety of periods to demonstrate the change in seasonal patterns.

4. Results

4.1 Annual trends

Discharge

Flow levels in the Massa and Gornera basins are insignificant in the months of November (11) through April (4) as there is very little meltwater generation. Any flow is drainage of meltwater stored from previous summer ablation flow and groundwater flow. Consistently in the Massa, months 5-9 (May-Sept) accounted for over 89% of annual discharge over the measured period between 2003 and 2011 (Table 4.1). Discharge in Gornera is only monitored by Grande Dixcence between May and October, as flow in the accumulation season is negligible and considered uneconomical. Table 4.1 shows the average percentage of annual flow for each month over the measured period. Outside of the ablation season (5-9), October has the greatest level of flow as dwindling radiation levels are able to generate some melt at the low elevations of glacial ice remaining free of snow cover after the ablation season and the release of subglacial water pockets which became charged during the ablation season. The comparative month to October at the beginning of the ablation season (April), shows very little discharge generated despite average incoming short-wave radiation levels being significantly higher than corresponding air temperatures. This is due to the presence of snow cover at the lowest elevation of the basin area.

Year-on-year analysis of the Massa ablation season (Table 4.2) shows that the month of greatest discharge commonly varies between July (7) and August (8). The always occurs after the summer solstice (21 June), indicating that the transient snow-line (TSL) continues to rise into the latter part of the ablation season exposing more low albedo glacial ice to solar radiation levels high enough to generate melting. Discharge rises sharply from May through to July before plateauing and falling steeply into month 9 (Figure 4.1). Overall means from annual Massa data, Table 4.2, show that 2003 was a year of much greater than average summer discharge and was followed by a lower than average discharge regime in 2004. 2003 has since been recognised as an exceptional year with a heat-wave dominating central Europe (IPCC, 2007). After 2003, 2009 is shown to be the year of greatest discharge levels in Massa.

Discharge levels in Massa and Gornera basins between days 0-120 and 270-365 (Figure 4.1) are so low that they can be regarded as insignificant for diurnal analysis and constant flow regime begins at day 120. Comparison of Table 4.2 and 4.3 and Figure 4.1 indicates Massa discharge levels are much greater than those experienced in the Gornera, due to the respective basin size and ice area. However, the patterns of low and high flows are very similar to one and other. Annual peak discharge at Gornera, between 2004 and 2008, ranges from June to August and is primarily linked to the draining of the Gornersee. July and August see the greatest levels of total discharge in the Gornera. Analysing the annual data from 2005 against one and other shows a high level of correlation between the sites (Correlation coefficient = 0.91) (Figure 4.2), indicating that changes driven in each basin are nearly identical.

Table 4.1 Percentage of total annual discharge in the Massa by month and for the period May through September ($Q_{5\text{-}9}$), for the year 2003 and 2011.

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	$Q_{5\text{-}9}$
2003	0.21	0.14	0.29	1.14	5.66	24.84	25.50	29.66	9.57	2.27	0.48	0.25	95.22
2004	0.21	0.20	0.31	0.76	3.89	15.00	24.97	29.24	17.32	5.01	2.65	0.45	90.42
2005	0.25	0.18	0.38	0.86	6.62	23.33	27.60	20.92	14.86	3.43	1.35	0.22	93.33
2006	0.14	0.11	0.14	0.61	6.13	19.22	36.05	14.15	15.32	6.46	1.28	0.37	90.88
2007	0.31	0.23	0.31	2.56	7.78	19.46	24.98	24.10	12.78	6.27	0.91	0.31	89.10
2008	0.21	0.23	0.28	0.42	7.45	20.85	25.82	26.79	13.66	3.09	0.83	0.38	94.57
2009	0.20	0.16	0.21	0.78	7.44	17.61	23.98	28.38	14.87	5.21	0.75	0.40	92.28
2010	0.27	0.15	0.22	0.94	5.78	18.87	35.08	23.13	11.00	3.56	0.69	0.31	93.88
2011	0.17	0.14	0.22	2.21	9.29	16.43	20.10	25.72	18.11	6.17	1.03	0.41	89.65
Ave.	0.22	0.17	0.26	1.14	6.67	19.51	27.12	24.68	14.17	4.61	1.11	0.34	92.15

Table 4.2 Mean discharge of the Massa by month and for the period May through September (Q_{5-9}) for the years 2003 through 2011 (m^3s^{-1})

	May	Jun	Jul	Aug	Sep	Q_{5-9}
2003	13.60	59.70	61.30	71.30	23.00	45.78
2004	6.20	23.90	39.80	46.60	27.60	28.82
2005	11.80	41.60	49.20	37.30	26.50	33.28
2006	11.00	34.50	64.70	25.40	27.50	32.62
2007	13.40	33.50	43.00	41.50	22.00	30.68
2008	13.80	38.60	47.80	49.60	25.30	35.02
2009	14.70	34.80	47.40	56.10	29.40	36.48
2010	9.77	31.90	59.30	39.10	18.60	31.73
2011	18.20	32.20	39.40	50.40	35.50	35.14
Mean 05-11	12.50	36.74	50.21	46.37	26.16	34.39

Table 4.3 Mean discharge of the Gornera by month and for the period May through September (Q_{5-9}) for the years 2004 through 2008 (m^3s^{-1})

	May	Jun	Jul	Aug	Sep	Q5-9
2004	1.96	7.21	13.28	12.86	11.15	9.29
2005	2.46	13.20	15.95	10.36	6.74	9.74
2006	3.33	10.67	21.97	9.57	7.97	10.70
2007	3.52	10.17	14.351	11.68	6.08	9.16
2008	4.33	9.99	15.00	14.50	7.35	10.24
Mean 04-08	3.12	10.25	16.11	11.79	7.86	9.83

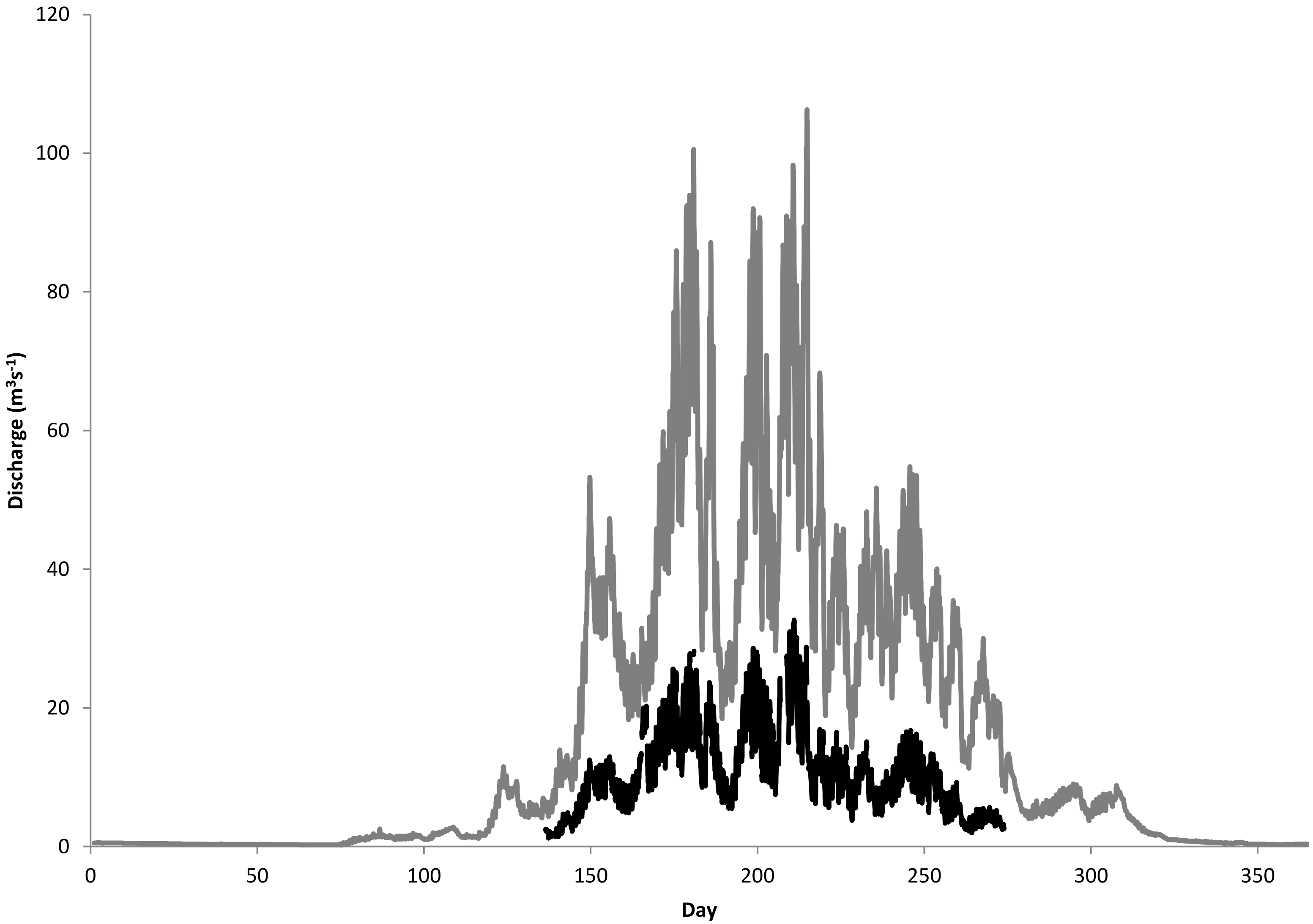

Figure 4.1 Annual hydrograph of the Gornera (black) and the Massa (grey), for 2005

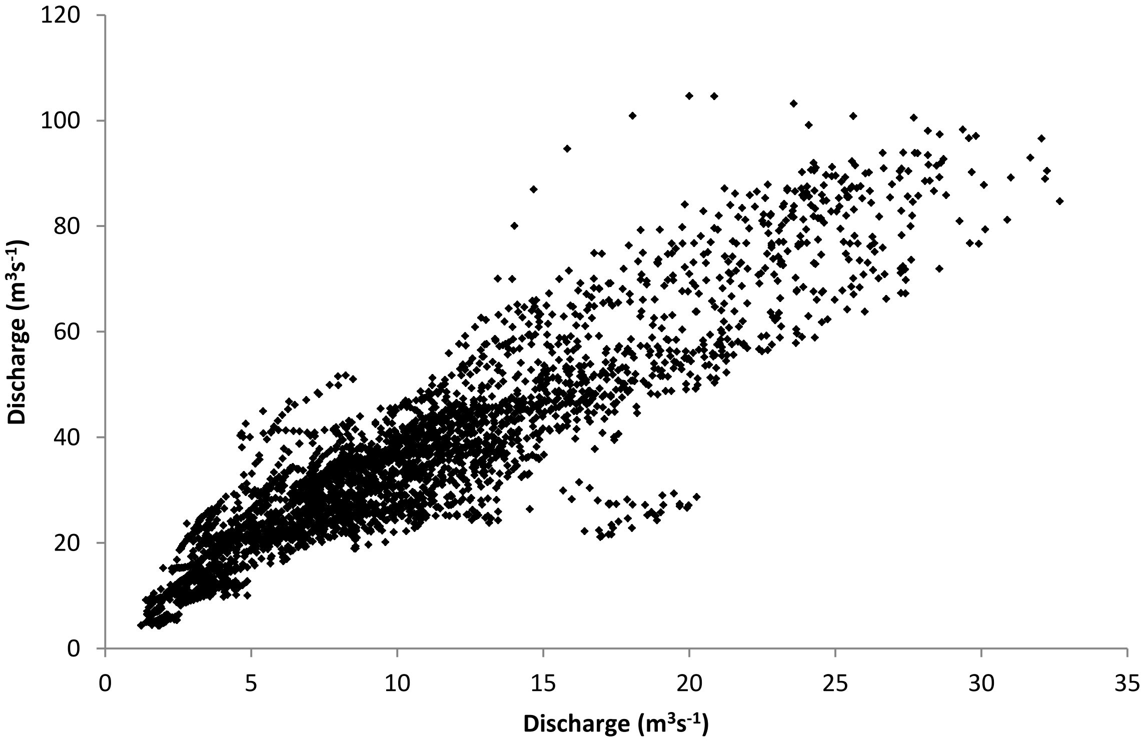

Figure 4.2 Plot of mean hourly discharges of the Gornera (x-axis) and the Massa (y-axis) for May through September 2007, r= 0.91.

Water temperature

Annual trends identified in Massa water temperatures show a previously undocumented spring maximum temperature period. Maximum water temperatures in the Massa are consistently found through April and May (Figure 4.3); Figure 4.3 is indicative of the study period. During this period residence times of snow melt, in the Massa, are long with low discharge and high solar radiation levels. Mean daily water temperature (Figure 4.3) highlights how the impacts of extreme fluctuations is reduced are that through June to September water temperature fluctuations are minimal in comparison with April, May and October, greatest range of readings are found in April with July showing the least fluctuation between maximum and minimum.

Through annual periods of high water temperatures, days where water temperatures were cooler are frequent, indicating a reliance of temperature change on incoming short-wave radiation levels early in the ablation season. The months with greatest variation in water temperature (Table 4.4) are those where radiation is at a substantial enough level to warm flow without generating large amounts of discharge, through icemelt. Early season is restricted by high albedo snow cover and late season fluctuation occurs as melt is reduced at higher elevations and discharge is low, increasing residence time and therefore contact with incoming short wave solar radiation is increased.

Table 4.4 Massa maximum and minimum monthly water temperature readings 2005 (°C).

Month	Max	Min	Range
Apr	4.30	0.20	4.10
May	3.12	0.80	2.32
Jun	1.62	0.63	0.99
Jul	1.56	0.75	0.81
Aug	1.72	0.80	0.92
Sep	1.74	0.76	0.98
Oct	2.18	0.59	1.59

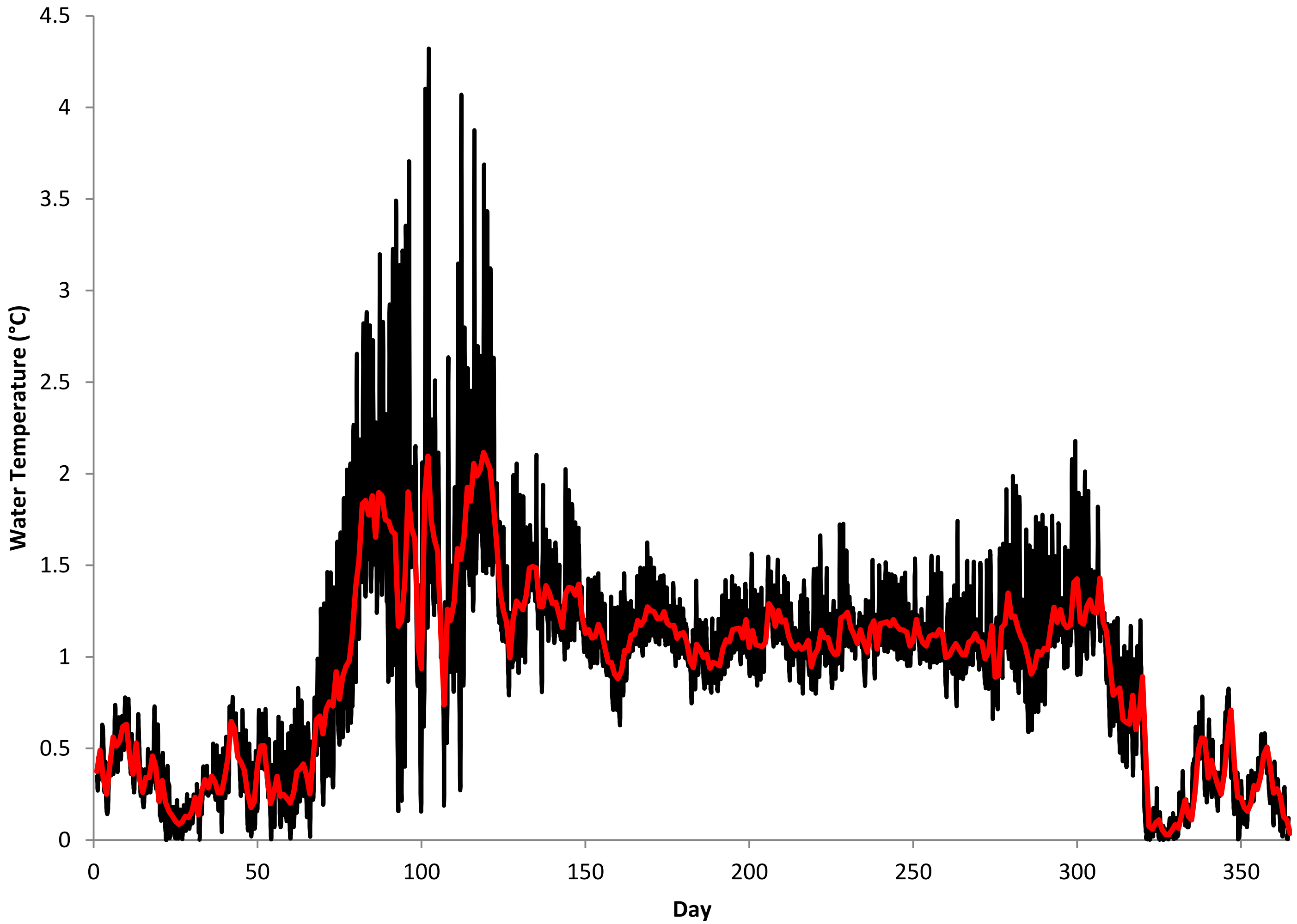

Figure 4.3- Annual water temperature variations of the Massa over 2005. Hourly values (black) and mean daily values (red). Distinct warm water pulse between days 75 (16 March and 125 (5 May).

Air temperature

Air temperatures in the Massa and Gornera basins fluctuate between summer maxima and winter minima. Annual air temperature regimes for Gornera and Massa basins over 2005 (Figure 4.4), indicate similar climatic patterns. Temperatures in both basins follow similar patterns with Massa having lower peaks and deeper minimums, due to its higher mean catchment elevation than Gornera.

Peak of air temperature in both basins in 2005, came on day 208 (28 July) with minimum temperature recorded on day 25 (25 January), giving an annual temperature range of 46°C (Figure 4.4). Positive air temperatures are required to generate any melt of snow and ice in basins and therefore the main focus should be placed upon the period of time where all recorded temperatures exceeded 0°C. Correlation of air temperatures is high between the two basins measured (Figure 4.5) indicating that both basins were subjected to similar climatic conditions.

Analysis of mean monthly air temperatures (Tables 4.5 and 4.6) shows strong trends between Massa and Gornera, with maximum annual temperatures occurring in July 2006 and July was proven to be the warmest month at both sites over the entire study period. Massa experiences higher ablation mean temperatures and lower winter means than Gornera, ablation averages are expected to higher. However, winter temperatures are anomalous as a site at higher altitude would be anticipated to experience lower winter temperatures. 2009 was the warmest year on record over the study period (where discharge, air and water temperatures were available) in terms of annual and ablation season mean temperatures at Massa (Table 4.5). Winters of 2006-07 and 2007-08 air temperatures at Gornera are abnormally high in January and February, with temperatures close to 0°C (Tables 4.5 and 4.6).

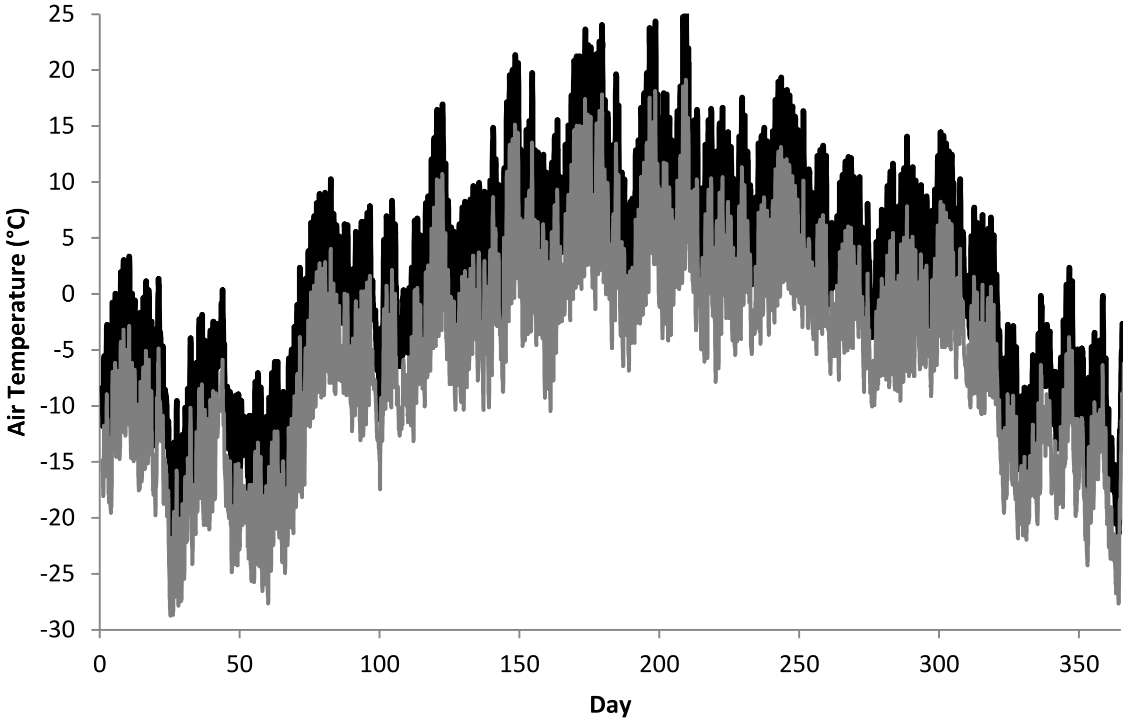

Figure 4.4 Annual variations of hourly air temperature at the Gornera (black) and the Massa (grey) over 2005. Corrected from Gornergrat and Sion Aeroport, respectively, using adiabatic lapse rate.

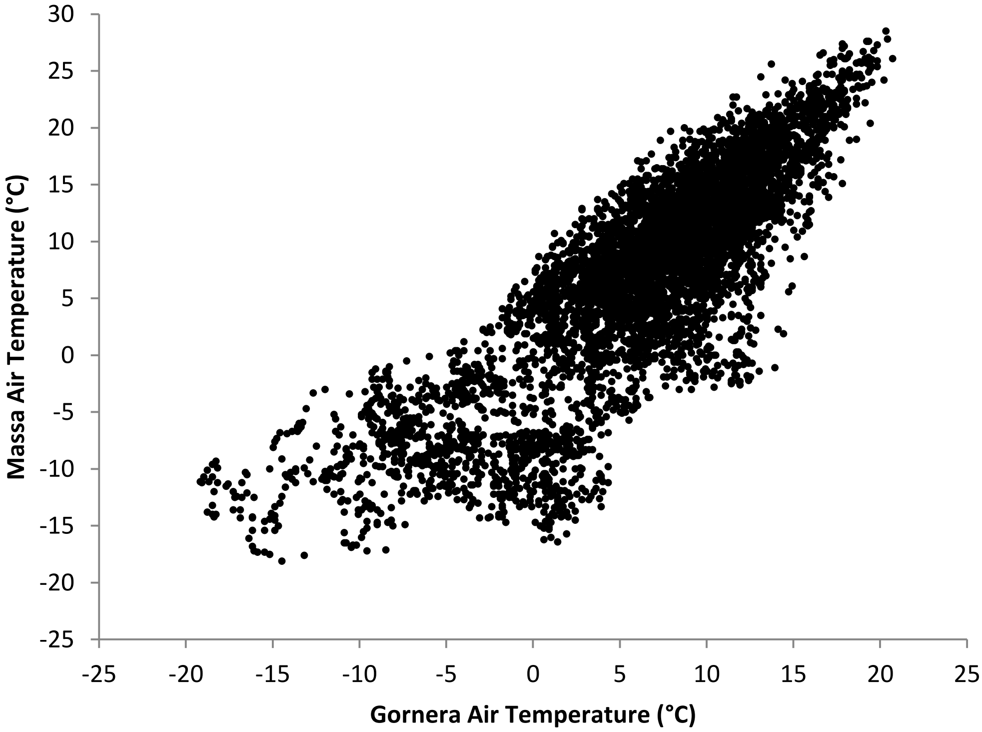

Figure 4.5 Plot of mean hourly air temperatures of the Gornera (x-axis) and the Massa (y-axis) for 2005. r= 0.84.

Table 4.5 Mean monthly and ablation season (T 5-9) air temperatures at the Massa gauging station between 2004 and 2011.

	2004	2005	2006	2007	2008	2009	2010	2011	Ave
Jan		-8.344	-8.403	-3.206	-3.764	-8.947	-7.653	-5.953	-6.610
Feb	-3.578	-7.557	-5.133	-1.887	-2.203	-5.201	-4.685	-3.338	-4.198
Mar	0.019	0.536	-1.734	0.985	0.150	-0.337	-0.125	1.726	0.152
Apr	4.870	4.378	4.595	9.038	3.960	6.158	5.619	7.951	5.821
May	8.074	9.588	8.636	9.242	10.704	10.187	7.760	9.979	9.271
Jun	12.484	14.038	13.251	12.388	12.993	12.242	12.682		12.868
Jul	13.712	14.055	17.127	12.983	13.905	14.304	15.552		14.520
Aug	13.546	11.452	10.617	12.377	13.415	15.395	12.261		12.723
Sep	10.265	10.454	11.630	8.321	8.484	10.600	8.609		9.766
Oct	5.838	4.786	6.560	3.737	4.489	4.126	4.104		4.806
Nov	-2.709	-2.671	0.192	-2.554	-1.756	0.528	-0.702		-1.382
Dec	-7.014	-8.269	-4.943	-6.057	-5.899	-5.189	-6.856		-6.318
Ave	5.046	3.537	4.366	4.614	4.540	4.489	3.880		4.285
T 5-9	11.616	11.917	12.252	11.062	11.900	12.546	11.373		11.830

Table 4.6 Mean monthly and ablation season (T 5-9) air temperatures at the Gornera gauging station between 2004 and 2009.

	2004	2005	2006	2007	2008	2009	Ave
Jan	-3.963		-3.108	0.344	0.646	-3.644	-1.945
Feb	-1.687		-3.933	-0.096	0.857	-5.165	-2.005
Mar	-2.171		-3.127	-2.078	-3.084	-2.145	-2.521
Apr	0.084		1.303	4.857	-0.117	1.349	1.495
May	3.706	6.027	4.985	6.313	5.574	7.335	5.657
Jun	8.937	10.531	9.786	8.824	9.336	8.777	9.365
Jul	10.747	11.016	13.869	10.452	11.215	11.951	11.542
Aug	11.655	9.436	7.782	10.534	11.770	13.461	10.773
Sep	9.871	8.902	10.824	7.671	7.246	9.604	9.020
Oct	6.371	7.356	8.597	5.887	6.037	5.816	6.678
Nov	0.815	-0.254	3.495	0.752	-0.372	2.092	1.088
Dec	-0.733	-3.910	1.097	-2.341	-2.791		-1.736
Ave	3.636		4.297	4.260	3.860	4.494	3.951
T 5-9	8.983	9.182	9.449	8.759	9.028	10.226	9.271

Inter-relationships between discharge, air temperature and water temperature

During periods where air temperatures frequently dip below and remain beneath 0°C, discharge (days 0-100) is not generated through any melting and so values are low, if not $0m^{3-}s^{1}$, any discharge that flows is sourced from groundwater . During periods where air temperatures are consistently above 0°C discharge is generated from melt, in the Gornera and the Massa this was found to be in May (day ~120) and sustained until the end of October (day ~300) (Figure 4.6). Peak discharges directly relate to peak air temperatures (Figure 4.6), and are seen to mimic air temperatures. Where temperatures dip considerably (days 185-190) in summer for a sustained period discharge levels are seen to fall in response.

Comparing readings of mean monthly discharge and mean monthly air temperatures between 2004 and 2011 at Massa (Tables 4.2 and 4.5), shows a relationship exists between the two variables during ablation seasons. Ablation season discharge means correlate strongly with air temperature means for the same periods, with the years of highest ablation season temperatures incurring the highest levels of mean discharge. Direct relationships between monthly variables are at their strongest in the closing months of ablation seasons (August and September), as during this period large portions of basins are snow-free and therefore large swaths of low-albedo glacial ice are exposed. Abnormal air temperatures in this period have a greater impact on discharge levels i.e. August 2006 was much cooler than mean (10.6°C, mean 12.7°C) and therefore discharge levels were effected ($25.4m^{3}s^{-1}$, mean $46.4m^{3}s^{-1}$). Early ablation season (May, June) impacts from air temperature on discharge are offset by blanket high-albedo snow cover, which restricts levels of meltwater generation.

Water temperatures when discharge levels are low, show much more variability than those where discharge is high and constant (Figure 4.7). Across the full range of collected data, water temperature variation reduced when discharge levels passed approximately 20% of the annual maximum discharge reading, variation returned in the closing months of the years as discharge levels fell below the dominating level. Variations were less marked at the end of the season than at the beginning of the ablation season (Table 4.4, Figure 4.7), due to lower levels of incoming short-wave radiation than in the early months (April and May). Water temperatures reach the highest levels in the Massa between days 75-125 (16 March-5 May). The length of warm water pulse, period of highest annual water temperatures, varies across the period studied and explanations of the warm water pulse are attempted in the following section. Water temperature variations reduce as discharge exceeds $25m^{3-}s^{1}$ (Figure 4.7), before again rising late in the year around month 10 before returning to regular winter levels.

During winter months when air temperatures rarely exceed 0°C, water temperatures remain close to 1°C (Figure 4.8) with peaks and troughs of air temperature being mimicked in water temperature data. This lack of change in temperature suggests that water is sourced from groundwater with a stable annual flow and annual temperature regime. Where air temperatures are seen to consistently remain above 0°C (day 70) water temperatures respond rapidly for a period, rising to a maximum of 4.3°C (day 101), before falling to a stable level, where water temperature fluctuations reduce markedly between days 125 and 300, this change occurs despite air temperatures rising to an annual peak and remaining consistently above 0°C. A sharp fall in water temperature occurs over days 106 to 109 caused by an equally steep decline in air temperature, this anomaly is likely caused by an external influence and ought to be deemed worthy of further investigation. Regression analyses between air and water temperatures (Figure 4.9), and, air temperature and discharge (Figure 4.10), indicate that there exists a lagged relationship between the two pairs of variables as r^2 values. However, it is not one strong enough for it to prove conclusively that the air temperature drives the dependant variables (discharge and water temperature). Where regression was plotted to prove a causal relationship between discharge and water temperature, no relationship was observed with r^2 values no stronger than 0.017.

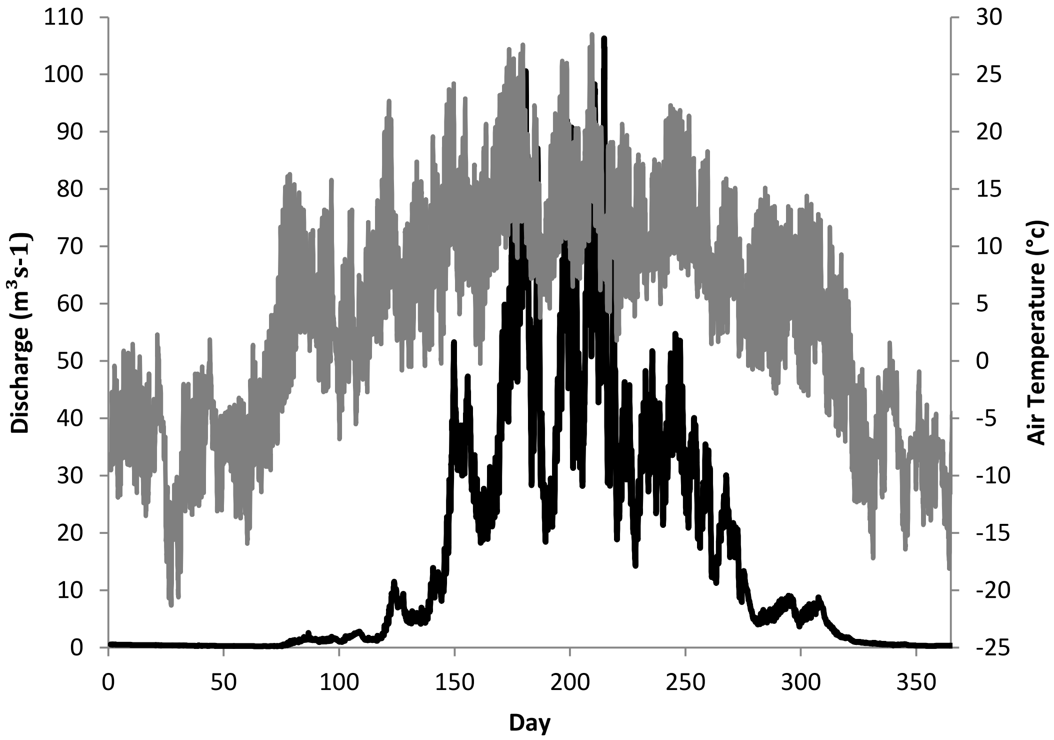

Figure 4.6 Annual variations of hourly air temperature (grey) and hourly discharge of Massa (black) for 2005

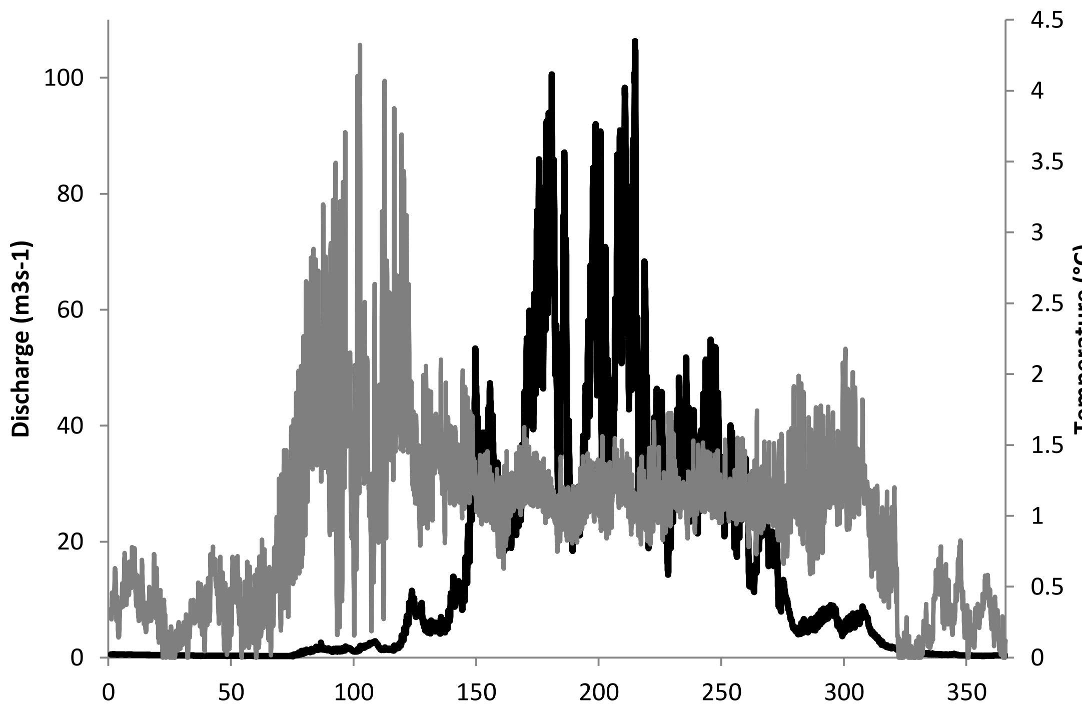

Figure 4.7 Annual variations of water temperature (black) and discharge (grey) pattern of Massa for 2005

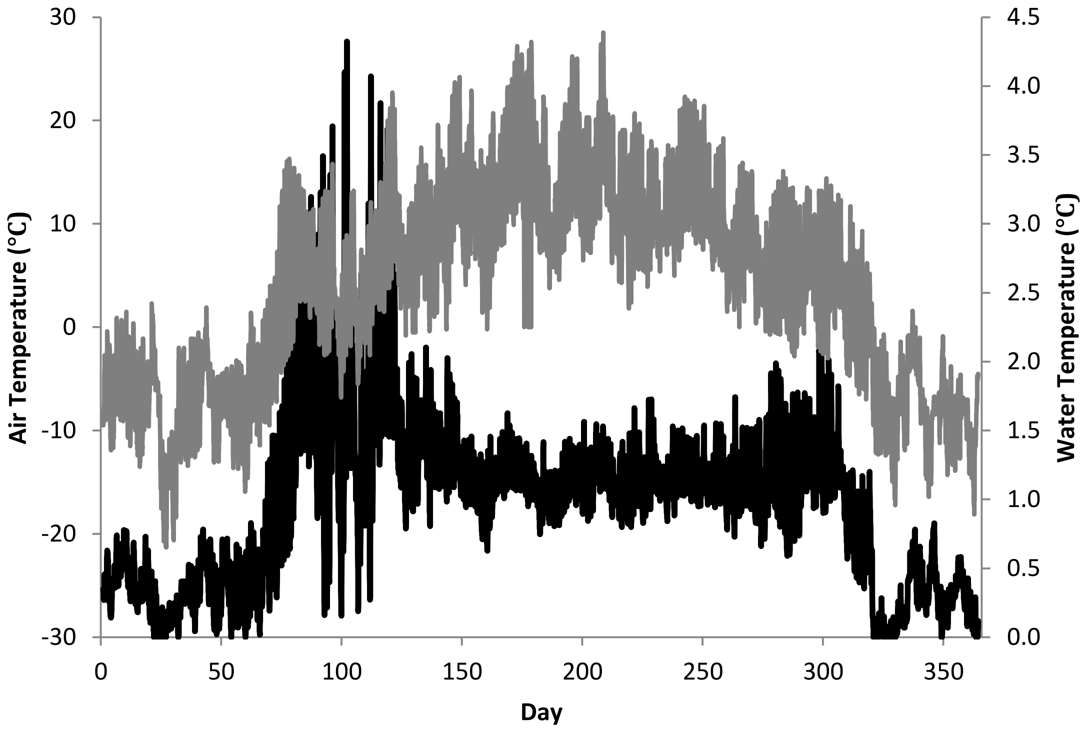

Figure 4.8 Annual air temperature (grey) and water temperature (black) of Massa basin for 2005

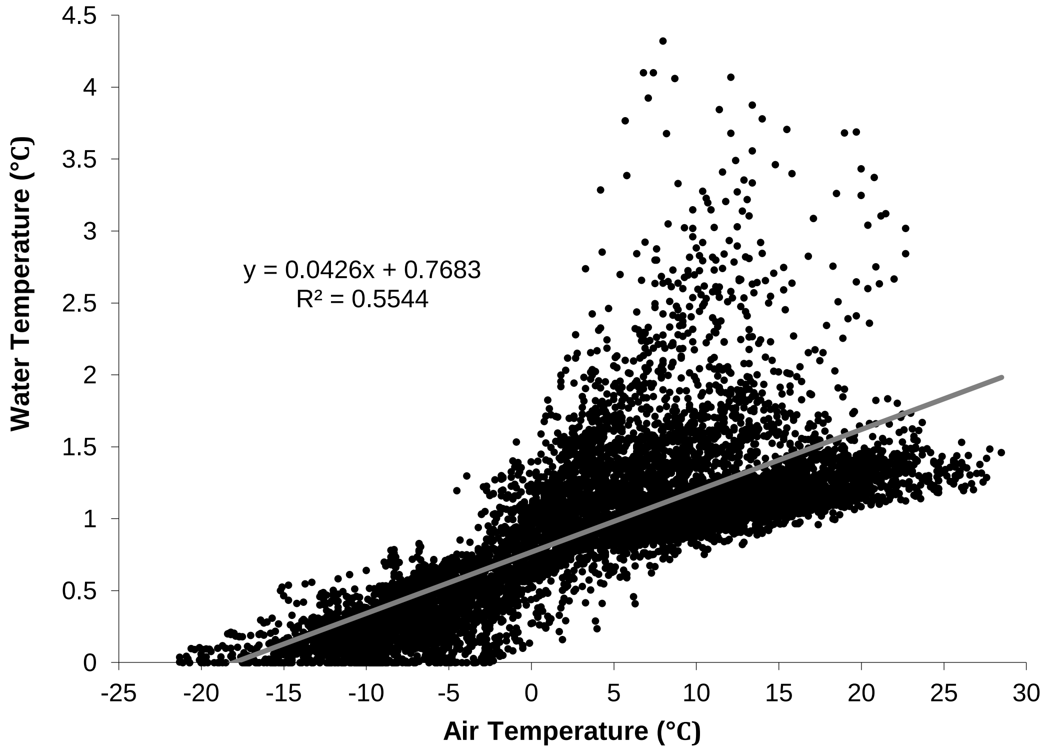

Figure 4.9 Annual regression plot of hourly air and water temperature in the Massa basin for 2005

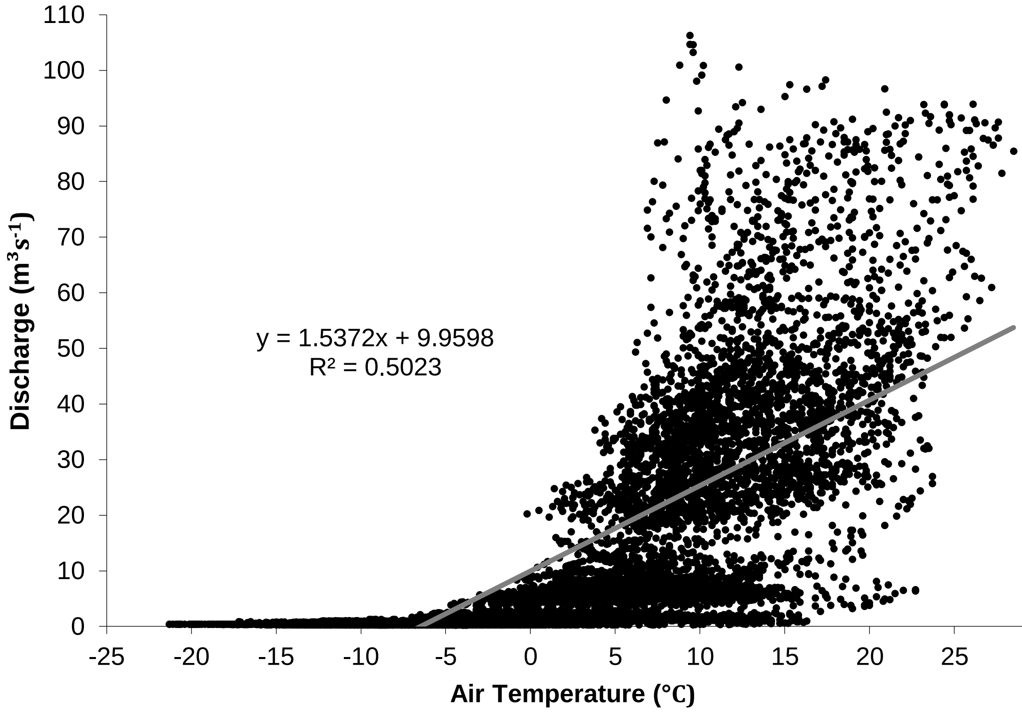

Figure 4.10 Annual regression plot of hourly air temperature and discharge in the Massa basin for 2005

4.2 Diurnal trends

Discharge

Diurnal discharge in both Massa and Gornera follows distinctive patterns of low and high discharges, discharges in Massa are more than three times greater than those observed in Gornera for the corresponding period (Figure 4.11), but fit well indicating that the two basins are driven by the same forces. Gornera and Massa basins experience minimum flow consistently at mid-morning (0800) with peak discharge occurring in late afternoon (1700), Massa reaches its peak one hour later than Gornera. Peak discharge is lagged due to the different segment lengths and the transit times of water sourced from snow and icemelt on the glacier, this lag is more pronounced during the early ablation season with minimum flow later in the morning and peak discharge occurring later at night. Massa has a threefold greater discharge than Gornera at peak discharge.

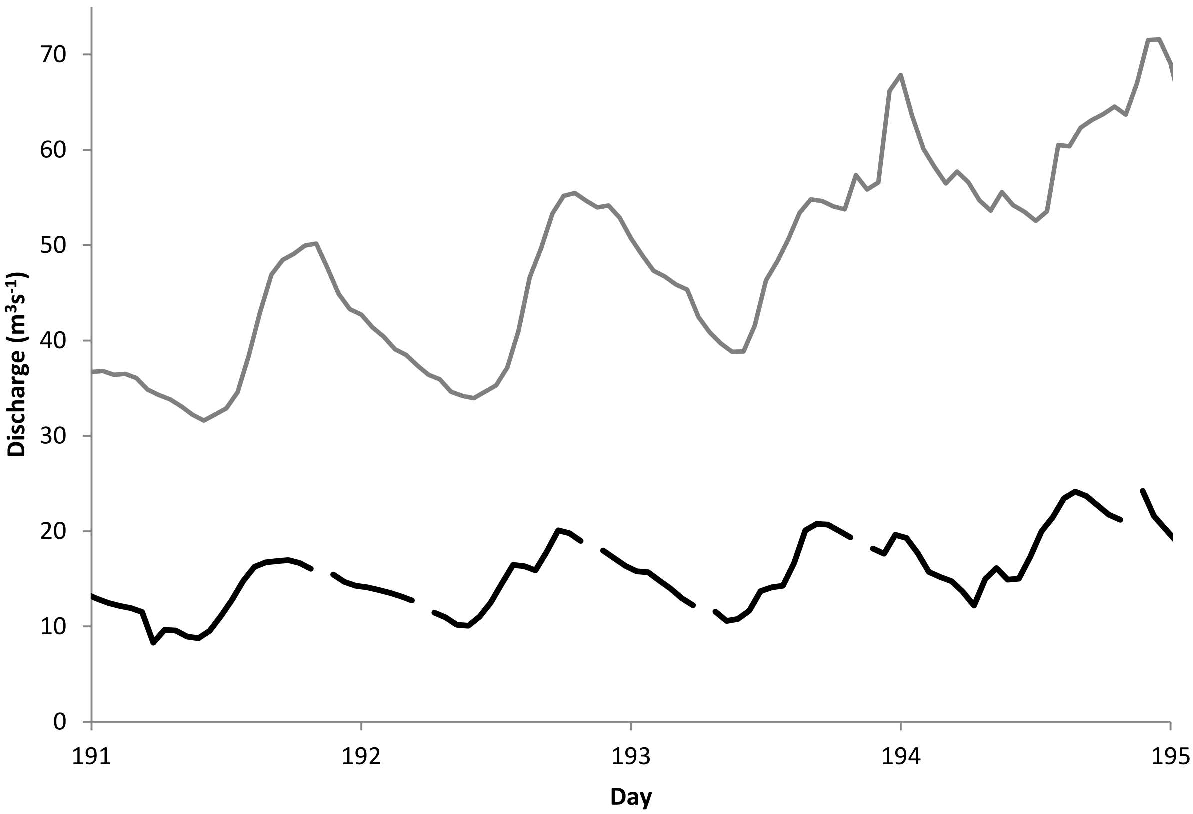

Figure 4.11 Diurnal variations of discharge in the Gornera (black) and the Massa (grey) for days 191-195(9-13 July), 2008. Gaps in the Gornera data indicate times at which the gauge was degravelled.

Water temperature

Diurnal water temperature data vary in both Massa and Gornera basins, showing that Massa water temperatures peak slightly later in the day than those at Gornera (Figure 4.12). The rivers exhibit a similar general pattern of water temperature regime with maxima and minima occurring at roughly the same time. Micro scale changes in temperature are mimicked at each site, with slight increases and decreases (0.1°C) representing that despite the areal distance between the catchments they experience very similar climatic conditions. Greater temperature change is observed in the Gornera, however maximum and minimum temperatures are higher in the Massa, primarily due to the extent of the segment and therefore the potential for temperature change is greater. Water temperature in Massa being warmer than Gornera by 0.5°C at their respective minimums but by only 0.2°C at the maximums, lower fluctuations in Massa are likely linked to the greater respective discharges, controlling temperature fluctuations. The rising limb of both rivers is steep and Gornera reaches the peak of temperature earlier in the day than Massa. The falling limbs of both streams are gentle in gradient. However, this is only the case as the temperatures shown occur late in the ablation season, at earlier and during the peak of the season the falling limb has a much steeper gradient.

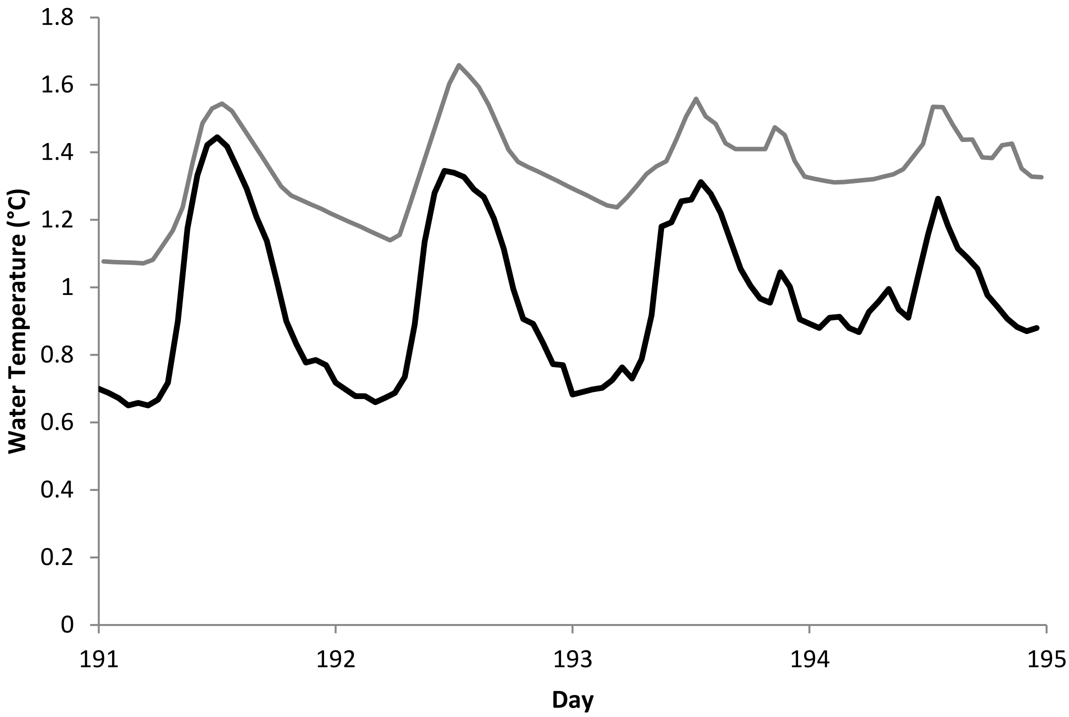

Figure 4.12 Diurnal variations of water temperature patterns in the Gornera (black) & the Massa (grey) days 191-195 (9- 13 July), 2008. Small scale changes in water temperature are observed in both basins.

Air temperature

Air temperatures follow a defined diurnal regime throughout the year with maximum temperatures occurring during the early afternoon between 1300 and 1400 at Gornera and Massa (Figure 4.13). Minimum temperatures are during the early hours of the morning, generally between 0100 and 0300, when incoming short-wave radiation is zero and reflected heat from other sources is lowest. Temperature increases coincide with the rising sun, the maximum daily temperature occur after the daily maximum of incoming short-wave radiation (1200) steeply increasing during clear summer days before steeply falling away after the daily maximum, as the sun descends toward the horizon.

The daily range of temperature is more pronounced in the Massa with higher daily maximums and lower daily minimums than the Gornera (Figure 4.13); the Massa basin experiences higher air temperature levels as the average elevation of the reach is less than that of Gornera and the basin is orientated North-South and is therefore exposed to more sunshine hours than the Gornera.

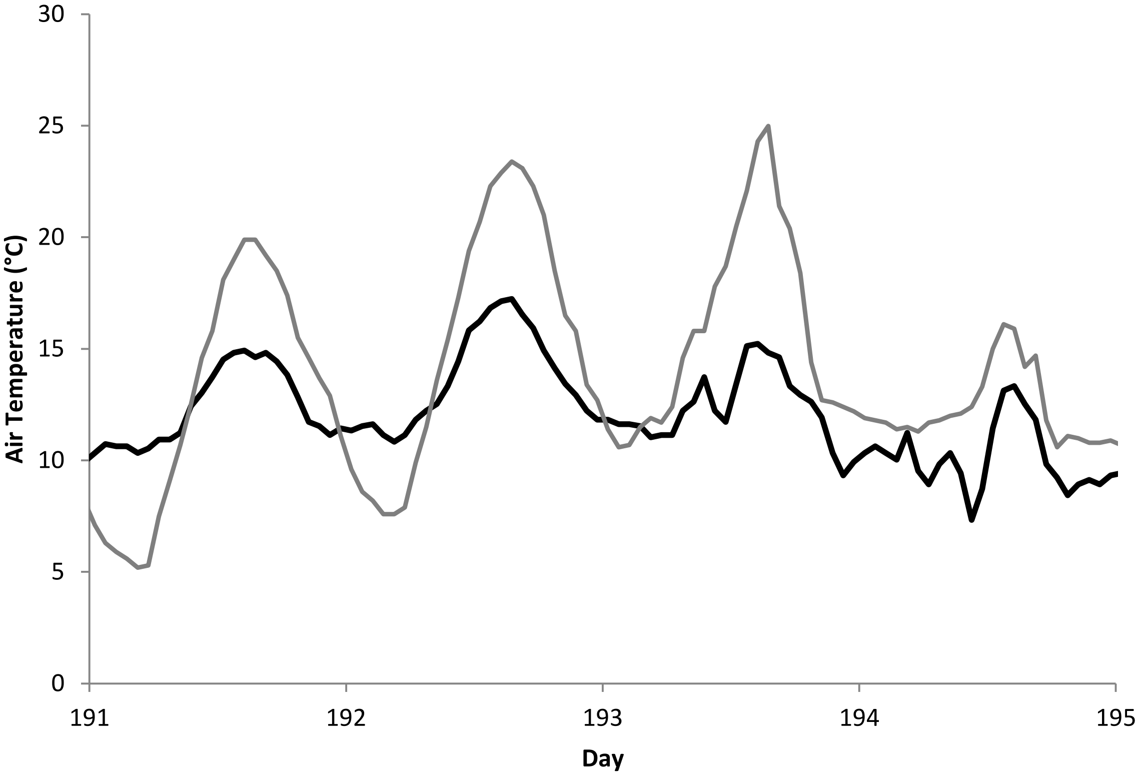

Figure 4.13 Diurnal variations of air temperatures at the Gornera (black) and the Massa (grey) basins for days 191-195 (9- 13 July), 2008

Inter-relationships between discharge, air temperature and water temperature

Water temperatures reach their maxima and minima before discharge (Figure 4.14). Both rising limbs of discharge and water temperature are steep, showing a rapid increase through the morning, temperatures reach the highest peak when discharge levels have fallen to lower levels than the previous day, the trend does not continue on day 249 with both variables staying relatively low suggesting a change in an external influencing factor.

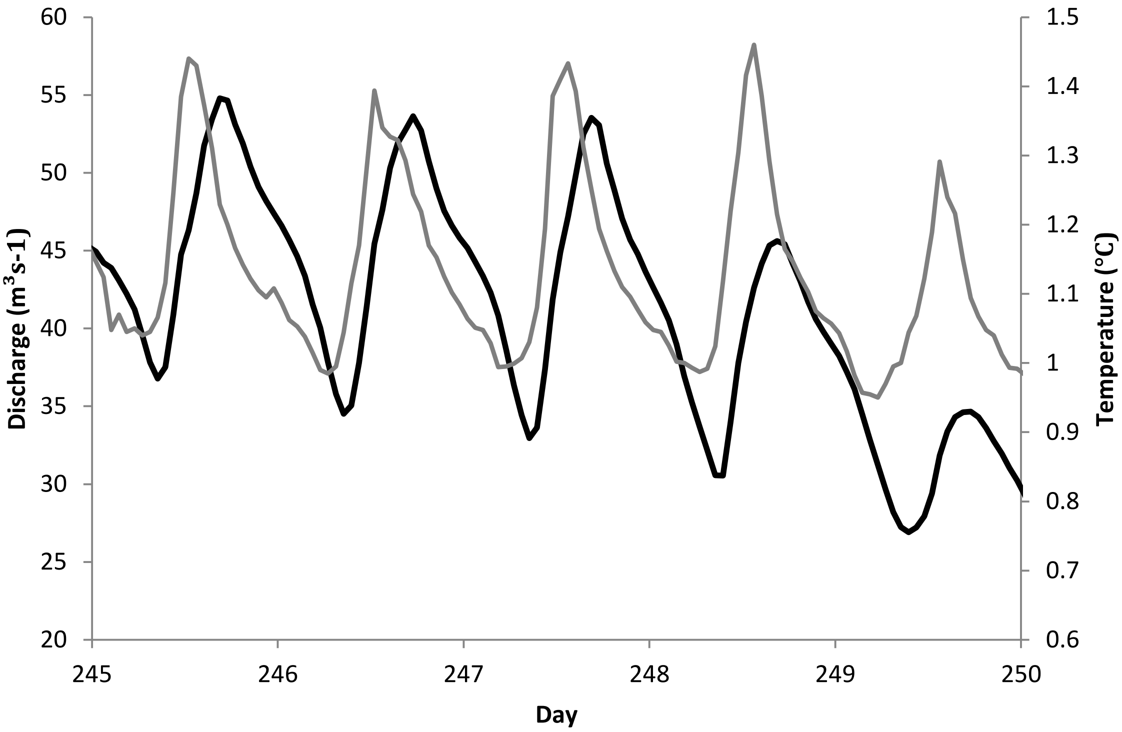

Figure 4.14- Diurnal variations of water temperature (grey) and discharge (black) of the Massa for days 245-250 (29 August- 4 September), 2005

Regression plots of discharge and water temperature in the Massa (R^2= 0.017) reveal no linear relationship between the two variables. However, hysteresis plots of discharge against water temperature (Figure 4.15) show that at different stages of the ablation season the peak of discharge is much later than that of water temperature, showing strong diurnal clockwise hysteresis. Water temperatures in August show a greater level of variability than those of September and increase greatly where discharge changes are low. September is the time where the sub-glacial drainage system is at its most efficient allowing the maximum amounts of meltwater to enter the pro-glacial stream. Increases in discharges therefore move in synchronicity with water temperature after 0800, when the sun rises until the peak of incoming short wave radiation at 1300. Discharge continues to increase until evening time as melting continues on the ice surface, even as incoming short wave radiation levels discharge continues to rise before decreasing at the close of day.

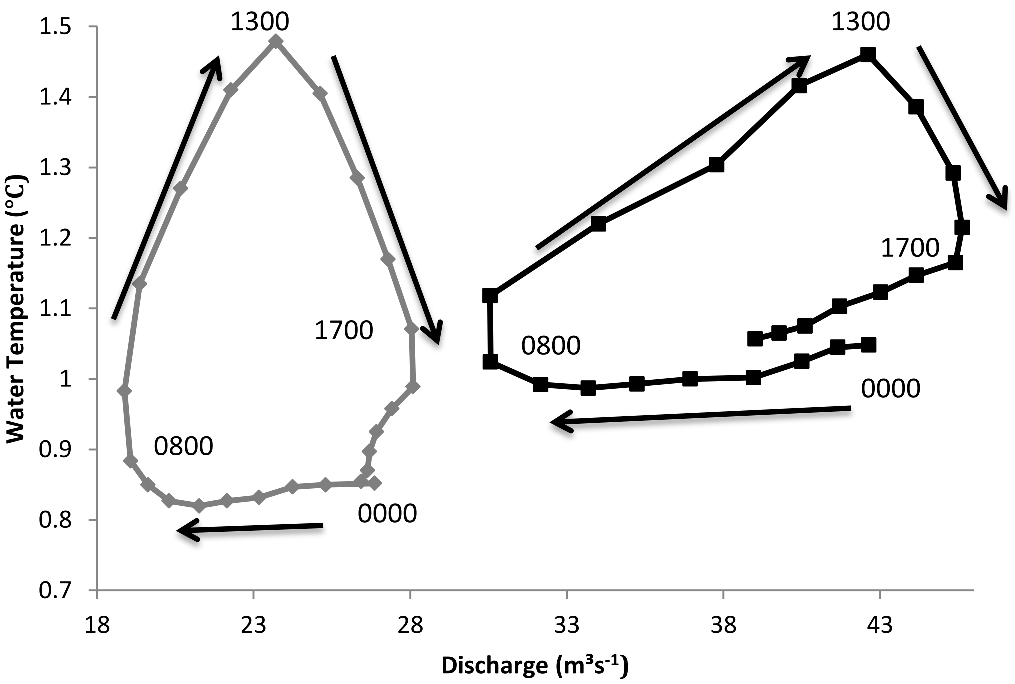

Figure 4.15 Hysteresis plots of diurnal relationship between discharge and water temperatures in Massa for single days in August (grey) and September (black) for 2005

Diurnal air temperature and discharge at Gornera follow similar patterns to water temperature and discharge (Figure 4.16), with a discernible lag of up to 2hours against discharge, maximum discharges occur later than maximum air temperatures and minima are also lagged by the same amount of time. The two variables rise in tandem with one and other with greatest peaks occurring together as well as the greatest minima, with disturbances in air temperature being mimicked in discharge data. Falling limbs of discharge are less steep as air temperature, long transit times for aliquots of water which do not enter into the main subglacial pathways cause them to be released slowly and creating an extended falling limb. Maximum air temperature and discharge both occur on day 174, with consistently high air temperatures from day 170 through to day 174 discharge levels increase.

Hysteresis plots of air temperature and discharge data show a cyclical diurnal pattern across the ablation season (Figure 4.17). Strong anti-clockwise hysteresis at different stages of an ablation season, early in season discharge levels respond slowly to changes in air temperature in contrast with later in season. When high albedo snow cover remains during June, large increases in air temperature elicit very little response from discharge indicating that very little flow generation from melting at this time. September, when the transient snowline is at a much greater elevation and the sub-glacial drainage network is developed indicates a rapid response to the rising of the sun from air temperatures (0800am) but a lag exists before discharge response (1000am) due to the length of the stream segment as flow is generated high up the catchment on the glacier. The lag is evident when at 1700 when air temperatures fall, discharge remains stable until 1900 before falling throughout the night where melting does not occur and flow is made up of runoff with the highest residence time within the glacier.

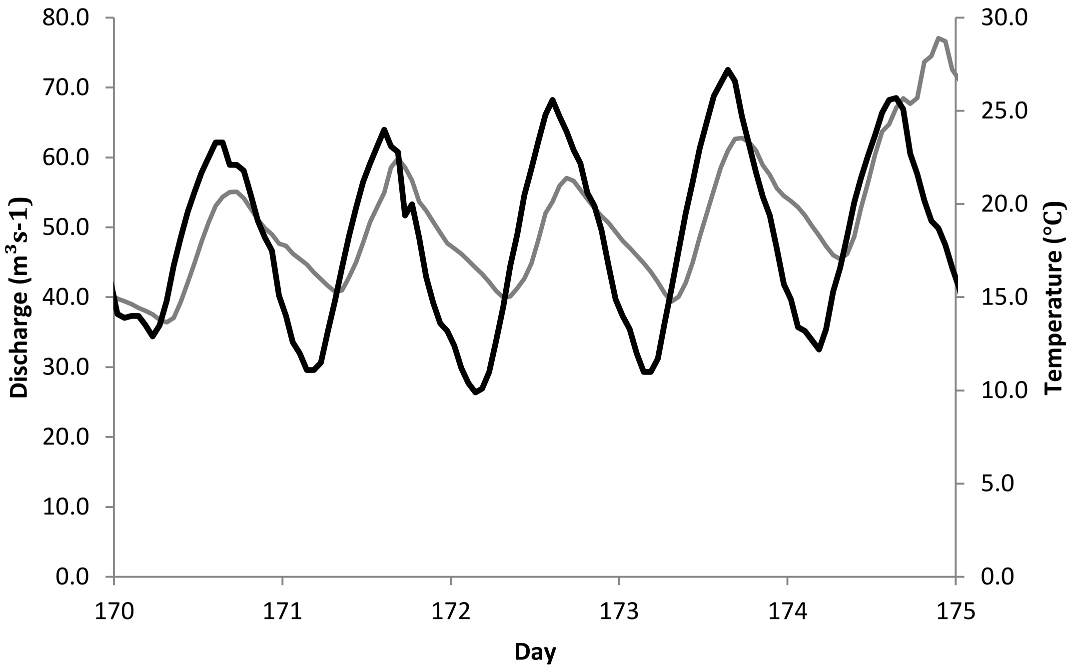

Figure 4.16 Diurnal variations of Gornera air temperature (black) and discharge of Gornera (grey) for days 170-175 (18- 23 June), 2005

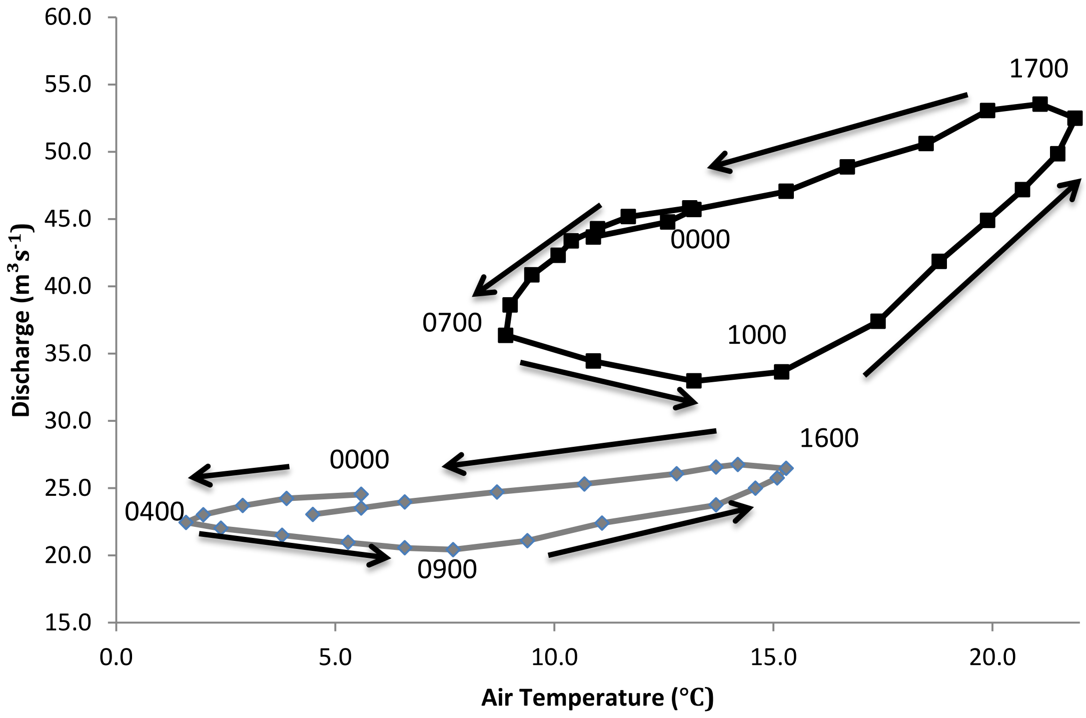

Figure 4.17 Hysteresis plots of diurnal relationship between air temperatures and discharge in Massa for single days in June (grey) and September (black) for 2005

Characteristic relationships between air and water temperatures occur on a daily scale throughout the ablation season, with two distinctive periods- early in the ablation season (Figure 4.18) and late in the ablation season (Figure 4.19); rising and falling limbs show similar gradients across periods with the peak of water temperature coming after that of air temperature. Early in the season water temperatures can be seen to increase to much greater levels than late in the season despite air temperatures being much greater for the latter period. Water temperatures reach their maxima on days 101 and 102. Day 104 shows that despite the increase in air temperature the water temperature was suppressed from reaching its previous maxima as air temperatures were lower. On day 104, it can also be seen that water temperature does not rise at the same rate as air temperature and after peaking returns to minima at a faster rate.

Water and air temperature data from different periods of time within the ablation season (Figures 4.18 and 4.19); indicate the progressive relationship between variables. Greatest variability in water temperature occurs where air temperatures are relatively low (Figure 4.18), late ablation season (Figure 4.19) shows part of the period described in 'Annual trends', where air temperatures are high but water temperature remains low and influenced by external factors, most notably discharge. Comparing the two time periods, indicates that early in the ablation season the greatest variability in range of both air and water temperatures.

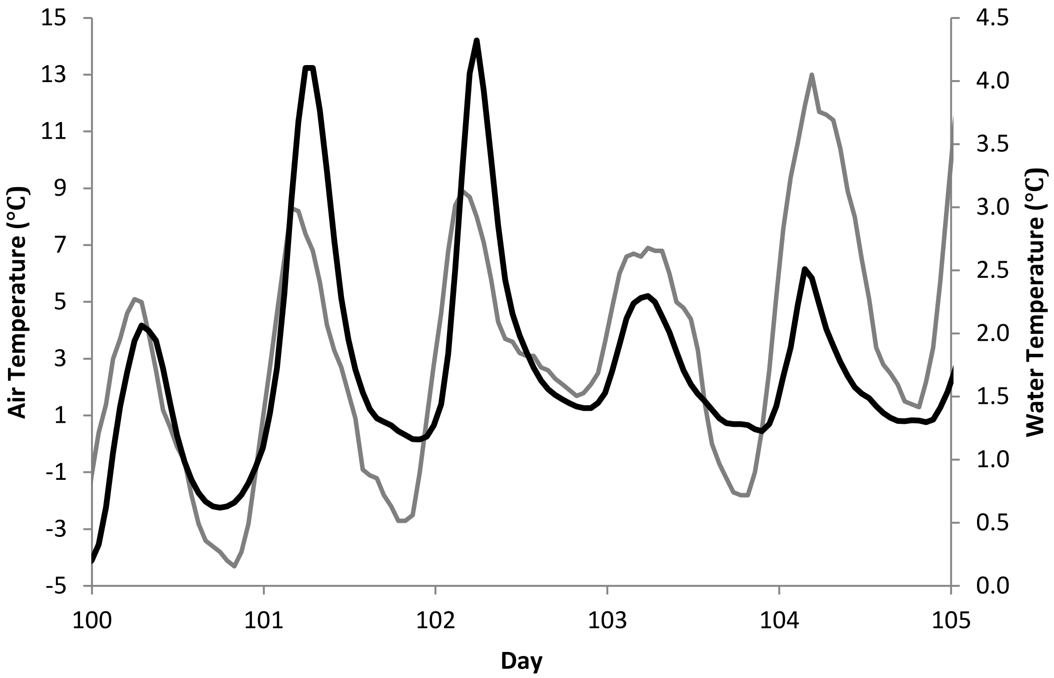

Figure 4.18 Diurnal variations of water temperature (black) and air temperature (grey) of Massa for days 100-105 (10-15 April), 2005

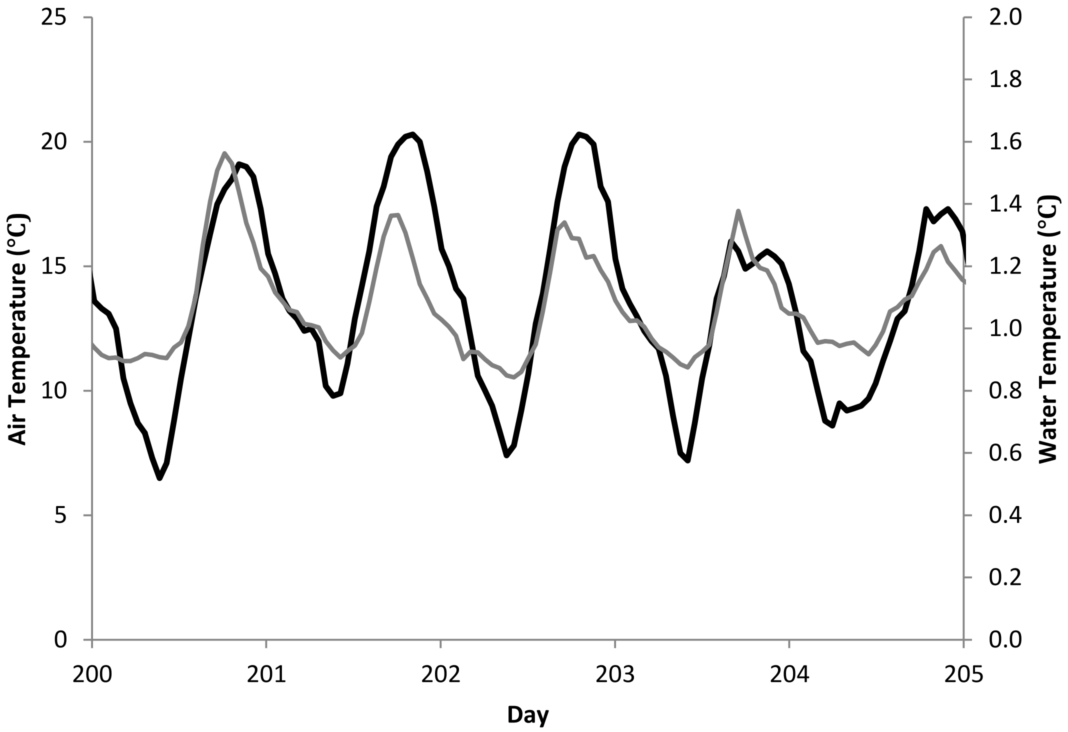

Figure 4.19 Diurnal variations of water temperature (black) and air temperature (grey) of Massa for days 200-205 (17-22 July), 2005

5. Discussion

Discharge for the Grosser Aletschgletscher system is three times greater than that of the Gornergletscher; as a result of the difference in size and scale of the two systems. However, despite the difference in total annual discharges from the respective basins, both basins receive approximately 90% of total discharge during the ablation season (months 5-9) from melting ice and snow.

The spring peak identified in the annual water temperature cycle in meltwaters issuing from glacier basins has not been previously identified. The work reported in this thesis is therefore an innovative contribution to the current literature. Focus of previous water temperature research in proglacial streams concentrated on paradoxical and contradictory relationships of air temperatures and incoming short-wave radiation levels being high yet water temperature remains low (Fellman *et al.*, 2013; Blaen *et al.*, 2012; Uehlinger *et al.*, 2003; Cadbury *et al.*, 2008).

At times at which water temperature peaks, air temperatures are not close to their annual maxima (Figure 4.8; Figure 5.1). However, when examining general annual patterns of short-wave radiation levels, they are much closer to annual peak explaining the large spike in water temperature. The lag which exists between peak annual air temperature and maximum short-wave radiation as the surface of the earth absorbs heat over time before releasing it back into the atmosphere creating a buffer to air temperature. This latent heat stored in the topography is a source of energy transferred into the water column through streambed interactions. Chikita *et al*. (2010) countered the argument that short-wave radiation or air temperature are responsible for the majority of water temperature increases when they stated that; at Gulkana Glacier, Alaska, the majority of heating was derived from the streambed interactions (38.2%) as opposed to short-wave radiation (32.1%). This however is unlikely to be the situation where glaciers occur in temperate regions when short-wave radiation levels are much greater. Glaciers at high latitudes receive much lower levels of incoming short-wave radiation than glaciers at lower latitudes.

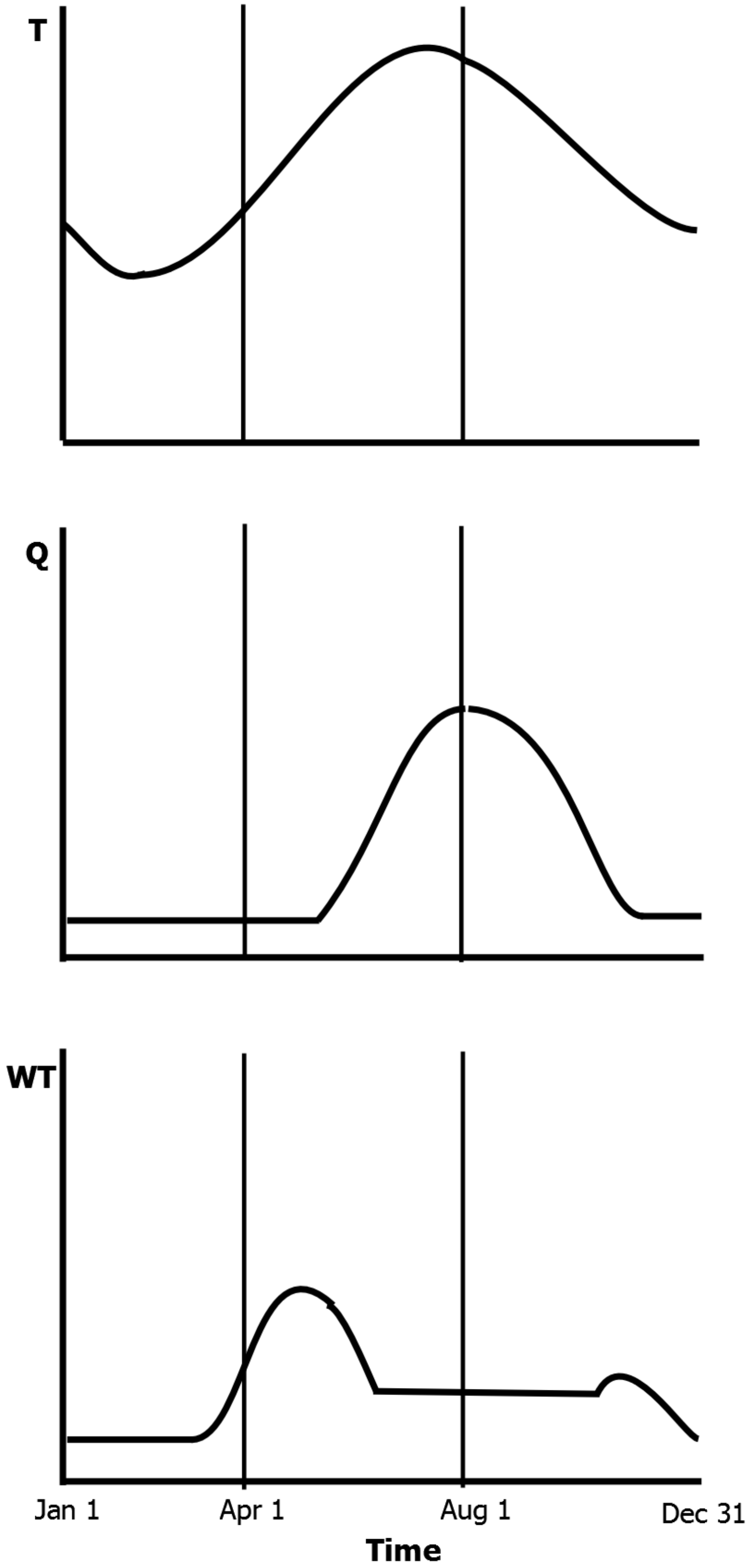

Figure 5.1 Schematic plot of annual patterns of air temperature (T), discharge (Q) and water temperature (WT) in rivers draining highly glacierised basins.

The timing and amplitude of the spring maximum water temperature observed in the Massa are heavily influenced by the snowfall of the previous accumulation season alongside high early summer short-wave radiation levels. Large accumulation of winter snow will increase the period where lower albedo glacial ice remains insulated and therefore delay the overall bulk of meltwater (Fleming, 2005). Short-wave radiation penetrates snow and ice to a depth of about 1m and 10m (Warren, 1982; Oke, 1987) and only 1–2% of global radiation (short-wave incoming radiation) penetrates into a snow cover (Konzelmann and Ohmura, 1995). Due to the exponential decline of transmitted radiation most of the energy is absorbed in the first few mm below the surface (Hock, 2005). This therefore extends the period within which high incoming short-wave radiation interacts with low levels of discharge creating an extended peak water temperature period. However, where snow accumulation is low, glacial ice will be exposed earlier creating a sustained pulse of discharge reducing high water temperature periods.

High water temperatures in the Massa occur from mid-March through to May (on occasions into June), where on average only 7.07% of total annual discharge is generated (Table 4.1), a proportion of melting which takes place is retained in the unsaturated firn layer due to re-freezing and capillary force (Tangborn *et al.*, 1975). This has been estimated as delaying up to 10% of annual snowmelt (Jansson *et al.,* 2003) entering the system at the time of generation before being delivered later in the season as snowpack melts rapidly. Data from the Massa for 2007 indicates that high water temperatures continued into June. Under close scrutiny there a dip in water temperature and coincides to dip in air temperature (Figure 5.2). Examination of data from the corresponding period at Stafel rain gauge, situated 1.6km from the Massa gauging station, shows that precipitation occurred on day 83 after period of days with cold air and water temperatures, suggesting that conditions were overcast and cloudy during this period and therefore little incoming short-wave radiation was available for stream heating. Snowfall was significant at low altitude, close to the gauge, and therefore would have been greater at higher elevations within the catchment creating a thick insulating layer of high albedo snow across the lowest extent of the glacier. This snowfall significantly sets back the generation of ice melt, and without this snowfall the onset of ice melt would begin around day 115 (26 April), earlier than any other measured annual cycle, but was delayed to day 155 (4 June).

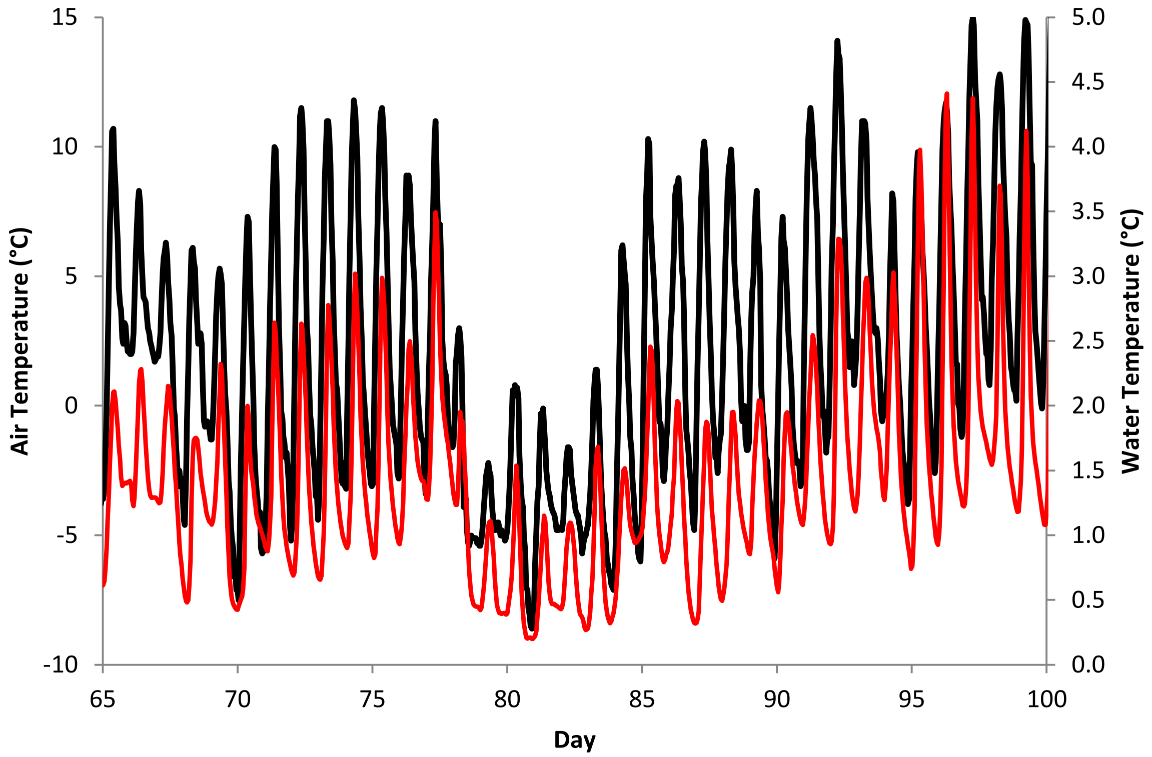

Figure 5.2 Massa water temperature (red) and air temperature (black) of days 65-100 (6 March- 10April), 2007. A period of low water and air temperatures is observed from days 73 through 85 (14 -26 March).

Table 5.1 Total annual discharge (Q5-9), mean ablation season water temperature (WT5-9) (°C) and air temperature (AT5-9) (°C) of Massa.

	Q Total	WT5-9	AT5-9
2004	122389.7	1.205	9.836
2005	132474.9	1.212	9.820
2006	N/A	1.645	10.356
2007	123964.6	1.460	9.723
2008	133534.1	1.442	9.724
2009	142926.8	1.415	10.441
2010	122383.1	1.310	9.521
Mean04-10	129612.2	1.384	9.917

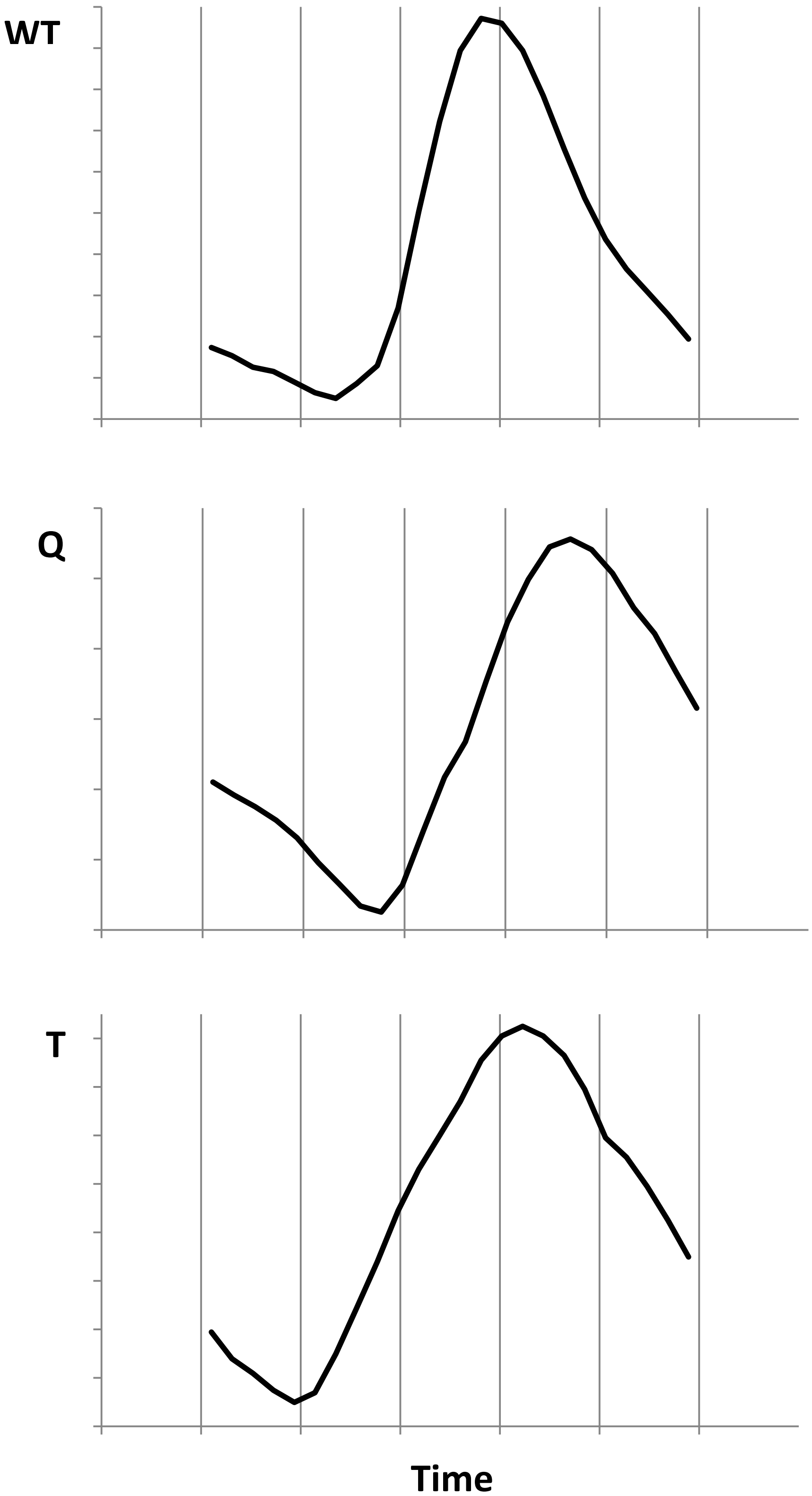

Figure 5.3 Diurnal pattern of air temperature (T), discharge (Q) and water temperature (WT) on day 210 (27 July), 2004 in the Massa. A clear distinguishable lag is present between the peaks of each variable.

Year-on-year analysis of ablation periods show patterns in average air and water temperatures and total discharge in Massa which indicate a complex series of relationships (Table 5.1). Large changes in annual discharge totals can be seen for very little alteration in average air temperatures, indicating external factors have a role in determining total discharge generated through snow and ice melt. During the warmest ablation period (2009), discharge is observed to also be at its highest however, water temperature is only slightly elevated above the average. This situation suggests a high proportion of discharge is generated through icemelt, which has a short residence time, reducing any impact from environmental interactions (Collins and Taylor, 1990). This contrasts highly with the situation of 2006, where air and water temperatures are higher than average. Without the availability of discharge data any relationships must be inferred. However, with a degree of confidence it can be suggested with high air temperatures and water temperatures, there is a greater component of snowmelt, which has a greater residence time than icemelt as previously explained allowing for greater environmental interactions.

5.1 Controls on Water Temperature

In order to understand Alpine stream water temperature dynamics the relevant controlling factors must be identified and their impacts fully understood. It is clear from results on both basins that multiple factors control the temperatures of waters issued from large Alpine valley glaciers (Figure 5.4), there are few primary controls dominating the majority of temperature change and many secondary controls which affect the primary controls, these factors are found to have greater effects across a wide variety of basins and differ throughout related literature.

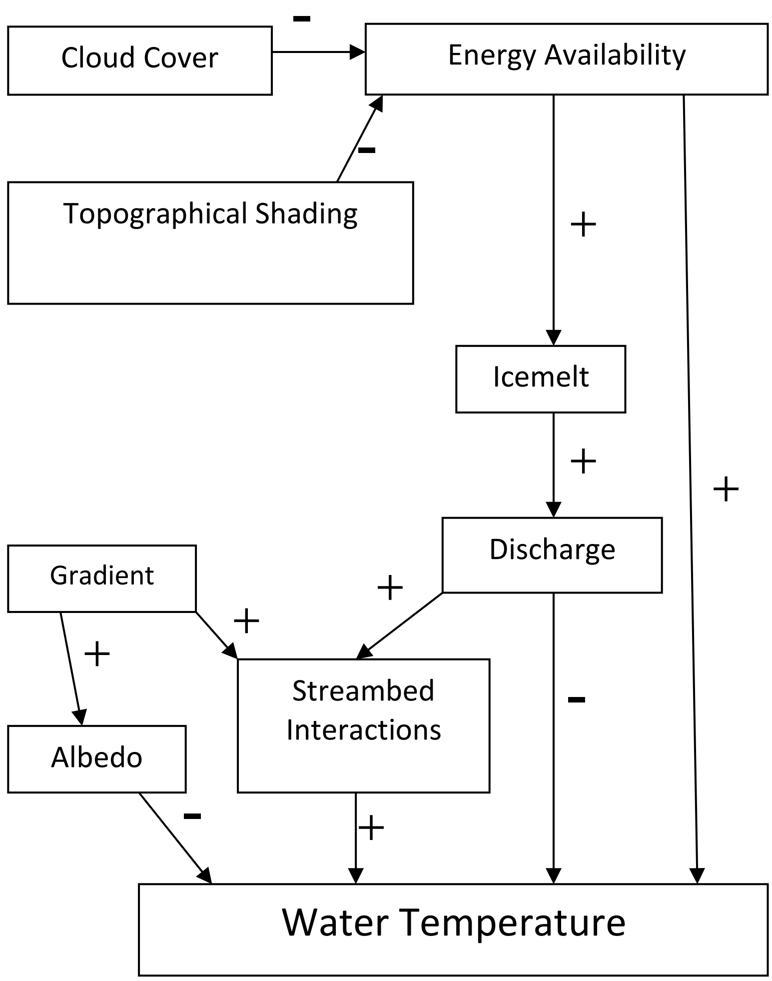

Figure 5.4 Plot of schematic flow chart showing the relationships between factors which affect water temperature in a river draining from a highly-glacierised Alpine basin.

5.1.1 Primary controls

Approximately 90% of total annual discharge in the Massa is in the ablation season (months 5-9), with the vast majority of flow sourced from the melting of snow and glacial ice (precipitation and groundwater account for a small proportion of summer flow) (Table 4.1), these waters enter the proglacial system at a temperature close to 0.1°C. Heating/ cooling rates are at their most pronounced in hydrological systems where there exists a wide gulf between water and air temperatures, it being the case that the greater the difference between the two then the greater the cooling or heating effect upon the water (Mohseni and Stefan, 1999).

Discharge is shown to be the most dominating factor for controlling stream temperature variations in the Massa on a seasonal and diurnal basis, when flow levels reach approximately 20% of their annual maximum flows and remain above the threshold over the ablation season. A tipping point is reached where by, when discharge is dominated by ice melt high variations are replaced by a consistent diurnal pattern of maximum and minimum temperatures. Increased stream velocity reduces the time available for environmental interactions to take place on aliquots of flow, and despite increases in streambed interactions they are offset by the sheer volume of flow. The same diurnal pattern is observed in Gornera during ablation seasons indicating that discharge controls water temperature variations. Incoming short-wave radiation levels and air temperature in the ablation season are seen to have much less of an impact upon water temperature change despite being higher than when the water temperature maximum occurs.

The use of hysteresis allows for clear demonstration of a cyclical diurnal pattern to show the relationship between air temperature and discharge (Figure 4.17) as well as discharge and water temperature (Figure 4.15). These cyclical patterns occur throughout the ablation season and are independent of scale, giving the same pattern from low levels of discharge and temperature through to high. Early in the ablation season discharge reacts very little to large scale changes in air temperature, signifying that high albedo snow remains across the majority of the glacier preventing melting, similar levels of air temperature change but at higher temperatures later in the ablation season (September) generate much greater levels of meltwater. Water temperature peaks close to short-wave radiation maxima (Figure 4.18) throughout the ablation season; however greater changes in discharge are seen later in season (September), when transient snowline is at a higher elevation exposing a larger area of glacial ice. Little temperature change in flow is observed during the night indicating that a constant force is responsible for maintaining water temperature level. This is likely to be derived from streambed interactions and bed friction (Chikita *et al.*, 2010). Blaen *et al* (2012) used hysteresis to illustrate the same effect, showing that water temperatures respond rapidly to changes in the atmosphere. However, discharge changes are lagged due to the generation of icemelt and subsequent transit of waters to the gauge.

Water temperature reaches its maximum at a time when air temperature is far from its annual peak. However, general patterns of short-wave radiation levels indicate that is relatively high explaining the large spike in water temperature (Brown *et al.*, 2006b). The lag which exists between peak annual air temperature and maximum short-wave radiation occurs as, the surface of the earth absorbs heat over time before releasing it back into the atmosphere creating a buffer to air temperature. This latent heat stored in the topography is a source of energy transferred into the water column through streambed interactions. Chikita *et al*. (2010) countered the argument that short-wave radiation or air temperature are responsible for the majority of water temperature increases when they stated that; at Gulkana Glacier, Alaska, the majority of heating was derived from the streambed (38.2%) as opposed to short-wave radiation (32.1%). Incoming short-wave radiation levels are reduced at high latitudes, and therefore this pattern will not exist in temperate regions where short-wave radiation levels are much greater (Hock, 2005).

Frictional heating, along the streambed and river banks, is not the overriding control of discharge in either the Massa or the Gornera, but it is considered to be a primary heat source alongside discharge levels, incoming short-wave radiation and air temperature (Moore and Richards, 2011). Streambed interactions are most prevalent in streams with a steep gradient, reflecting the high level of bed friction (Chikita *et al*., 2010); both the Massa and Gornera follow steep gradients. As discharge increases so does wetted perimeter, therefore raising the amount of direct contact between flow and the streambed and increasing the level of frictional heating. In large single channel river systems, like Gornera and Massa, as discharge increases the stream width remains at the same level with depth increasing greatly. At high discharge, frictional heating levels are greater. However, the amount of water to be heated is also greater and therefore increases in energy input are offset by rises in discharges and the rate of frictional heating remains close to a steady state level. Frictional heating at high discharge in Massa will be greater than at Gornera, due to the greater levels of discharge. Gornera will see a greater level of frictional heating than Massa at comparable discharges, as the steeper gradient increases the amount of energy generated per unit area (Chikita *et al.*, 2010).

Through observing water temperature patterns at high discharge during the night, where incoming short-wave radiation is nil and air temperatures are low, the amount of heating generated through frictional heating can be estimated from observed minimum temperatures. Where inputs from other sources are minimal, the majority of change will come from the streambed as the level of interaction with high discharge is the only heat source to remain high. Water emerging from the glacier portal at temperatures close to 0.1°C, will have been subjected to frictional heating where it has passed in contact with the glacier sole, however, any heating will be off set as waters are replenished by newly derived ice melt and water from englacial channels, close to the portal (Rothlisberger and Lang, 1987), which have only been in contact with ice so frictional heating will be close to nil.

5.1.2 Secondary controls

The albedo of flow has been recognised as being a key cause for the reflection of incoming short-wave solar radiation (Richards and Moore, 2011). Discharge which is dominated primarily or solely by snowmelt will culminate in low albedo even during periods of high discharge, as little to no suspended sediment is carried within the waters and back-scatter of solar radiation can only come from the presence of whitetop waves. During the ablation season where discharge is dominated by ice melt; flow is sediment laden and turbulent even at low discharge (Collins, 1979c), therefore, increasing albedo of the stream. Stream gradients greater than 10% add surface albedo through the creation of whitetop waves at high discharge according to Richards and Moore (2011). Gornera has a gradient of 20% actual fall and Massa 13% actual fall respectively (Table 5.1), so this impact can be attributed to reducing total energy flux at high discharges during the periods of highest air temperature and greatest incoming solar radiation. Albedo will be at its maximum relative level in the Gornera due to the gradient despite turbidity levels, which can affect the amount of shortwave radiation reflected back into the atmosphere, being greater in the Massa. Richards and Moore (2011) found that despite enhanced levels of suspended sediment concentrations, it could not account for the apparent dependence of albedo on discharge.

Precipitation effects on stream water temperature are not fully examined in this study, however an increase in stream water temperature will occur when rainfall events are responsible for a large enough percentage of total discharge to affect the overall ice-melt dominated system. As cloud cover is required for precipitation to occur, direct solar radiation is not able to impact upon a glacier surface and therefore the generation of meltwater is reduced significantly (Collins 1998). Peak of discharge is lagged to air temperature due to the transit time from the point of origin on the glacier to the pro-glacial stream, so in order for precipitation to have a marked impact upon water temperature cloud cover must be consistent for a long period before a large input of precipitation into the non-glacial area of the basin.

Collins (1998) analysed the response of the Gornera to summer precipitation events and their effects upon the diurnal discharge regime. It was found that the quantity of water contributing to rain-induced increases in flow is related to the total amount of rain during a storm, and to the partial area of basin which is both snow-free and receiving liquid precipitation. Storm events in the early ablation season are likely to produce high-albedo snowfall, as shown at the head of this section creating an extended warm water pulse (Figure 5.1), whereas later in the season a greater proportion of liquid precipitation would enter the system as the transient snowline and 0°C isotherm increase in elevation. The effects of precipitation are felt greatest in basins with low percentage glacierisation, where low ice runoff levels mean the cold water pulse from glaciers in a basin is easily offset by precipitation across the large swaths of ice-free area and rendered ineffective in controlling water temperature. Brown and Hannah (2007) investigated a Pyrenean basin where icemelt was only a minor contributor to flow and found that water temperature decreased as a result of precipitation input and incoming short-wave radiation reduction. In the Brown and Hannah (2007) study, glaciers account for a very small percentage of basin cover (0.07km^2) and as such the discharge regime is dominated by groundwater and contributions of seasonal snowmelt, summer precipitation and icemelt.

Topographic shading restricted direct solar incidence at Gornera when the sun was low in the sky (Figure 5.5). During the early and late ablation season, the topography on the southern side of the stream is steep enough to cast the stream in shade for long periods, the glacier however remains in direct contact with incoming short-wave radiation, generating melting and therefore discharge. At the onset of the ablation season when topography shades the stream, meltwater generation is low due to the low altitude of the transient snowline, yet when the same occurs at the end of the ablation season the transient snowline is at its greatest elevation with the maximum annual area of ice bare of snow (Collins, 1984). Discharge levels are therefore seen to be greater at the end of the ablation season than at the beginning; despite incoming short-wave radiation levels being lower the level of the transient snowline and 0°C isotherm are at their greatest elevation allowing low incoming short-wave radiation levels to generate significant discharges.

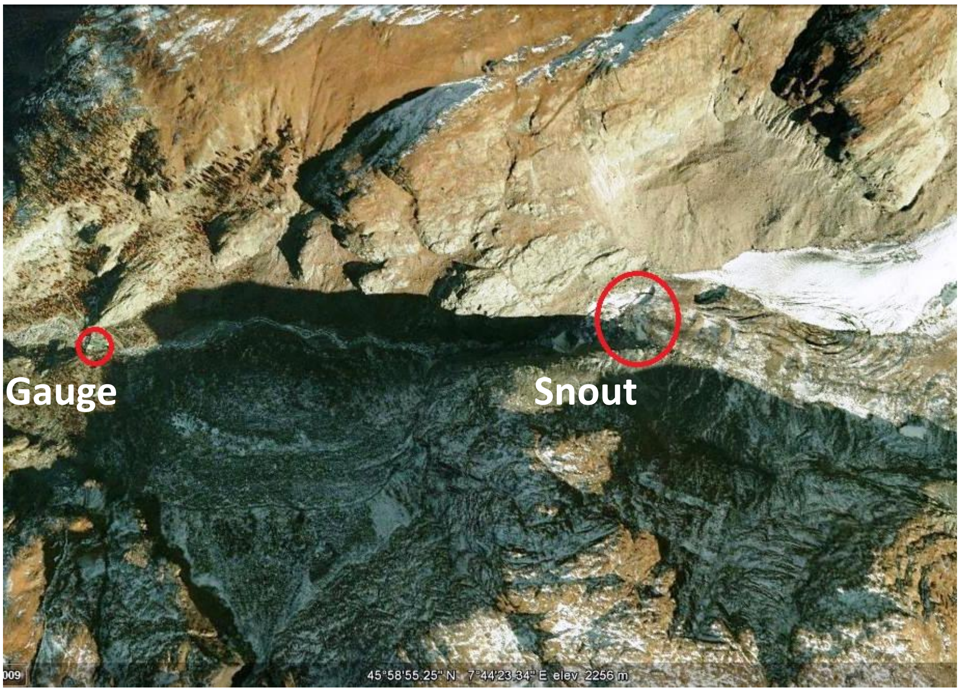

Figure 5.5 Satellite imagery showing topographical shading of the Gornera in October 2009 (Google Earth, 2013)

5.2 Warming Rates

The rates of stream warming observed over the study period at both Massa and Gornera were lower than those previously observed by Cadbury *et al.* (2008), Uehlinger *et al.* (2003) and Brown *et al.* (2006a) in other Alpine basins (Table 5.2). A lack of warming at Gornera and Massa can be attributed to topographical shading and discharge quantities from icemelt. The previous studies focussed on basins with a maximum of 30% basin glacier cover. Brown *et al*. (2006a) observed warming rates of 7°C km^{-1} in a steep proglacial stream in the French Pyrenees, however, catchment glacier cover was below 5% and only basins with glacier cover over 10% are controlled by the balance between ablation and accumulation seasons (Fountain, Tangborn, 1985). In this study the majority of stream flow was generated by groundwater (discharge was 0.07 m^3s^{-1} 1km from the glacier portal) and at the measurement site multiple water sources had converged to compose the main flow. The results found by Uehlinger *et al*. (2003) were similar to those found in this study; both studies investigated waters issuing from glaciers in above treeline reaches in the Swiss Alps, with the exception that Uehlinger *et al.* investigated a glacial floodplain. However, discharge and channel gradient were much lower than observed at Massa and Gornera, therefore the similarities in the results can be attributed to the annual climate of the Swiss Alps. Stream temperature variations among different channels within the floodplain were higher than along the entire main channel, not just in the floodplain suggesting that the bulk of discharge offset these higher temperature inflows.

Where glacial cover is significant (>30%), water temperature heating rates are similar across different studies, showing a slight decrease where glacial cover is over 50%. The heating rates are seen to be similar across a variety of different elevation scales with the key difference represented by average discharge. It is clear therefore that higher velocity, due to increased discharge, is responsible for reducing the amount of interaction time between atmosphere, streambed and an aliquot of water. Fellman *et al*. (2013) stated that a glacier with 29% basin coverage resulted in low variability of stream temperature with ice melt insufficient to reduce stream water temperature however, discharge levels were sufficient to prevent increases with air temperature. Conducive to the statement that glaciers with greater than 30% basin glacier coverage control the level of proglacial stream temperature variations.

Fellman *et al.* (2013) investigated stream temperature responses on a number of rivers issuing from variably glacierized basins and correctly identified that water temperature decreased with increasing glacier coverage. However, distance downstream of study sites and warming rates are neglected, this is demonstrated by the data that show basins with second and third largest levels of glacierization are only slightly warmer than the most glacierized basin despite waters travelling a much greater distance downstream. Therefore, warming rates in these streams are lower than the most glacierized basin, contrary to the data stated in Table 5.2.

Table 5.2 Comparison of studies of Alpine streams

	Massa	Gornera	Cadbury et al. (2008)	Uehlinger et al. (2003)	Brown et al. (2005a)
Location	Berner Oberland, SWI	Pennine Alps, SWI	Rob Roy Glacier, NZ	Bernina Massif, SWI	Pyrenees, FR
Glacier cover	65%	55%	30%	30%	<5%
Elevation (m a.s.l)	1800-1446m	2315-2000m	800-400m	2064-1998m	2550-1850m
Channel gradient	13%	20%	12%	2%	70%
Warming during melt season	0.45°C Km^{-1}	0.5°C Km^{-1}	0.6°C Km^{-1}	0.6°C Km^{-1}	7°C Km^{-1}
Discharge	34.39 m^3s^{-1}	10.39 m^3s^{-1}	4m^3s^{-1}	2.8 m^3s^{-1}	0.07 m^3s^{-1}

5.3 Future implications of climate change

If global air temperatures continue to rise and glaciers continue to recede, it is important to anticipate the changing landscape of water temperatures issuing from Alpine glaciers. It can be suggested that as glaciers recede and take up a smaller percentage of basin area that they will begin to behave in a similar way to those at current levels in the modern day. However, it is important to understand that these glaciers would have to contend with greater annual air temperatures and therefore increased levels of melting contributing larger masses of meltwater to the proglacial environment.

As the atmosphere warms, large glaciers will reduce rapidly in volume creating a pattern of increased annual runoff (Figure 5.6) (Jansson *et al.*, 2003). This increase in annual runoff will lead to an enhanced control of discharge on water temperature variations during the core of an ablation season when air temperature and short-wave radiation levels are highest. However, in the long term when glacier volume is low, annual runoff will become dominated by snowmelt, precipitation and groundwater. Basin water temperatures will move away from being dominated by icemelt and be more representative of air and groundwater temperatures.

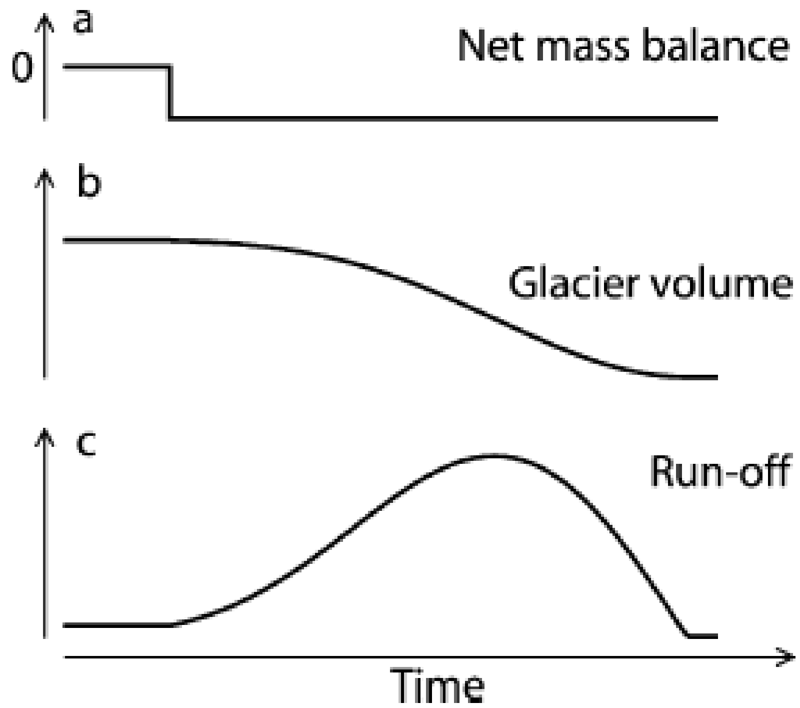

Figure 5.6 Schematic representation of the long-term effect of negative glacier net mass balance (a) on glacier volume (b) and annual glacier runoff (c). Volume response lags forcing due to the time required to remove ice by melting (Jansson *et al.*, 2003)

As glaciers recede the distance to the gauge inevitably increases (Brown *et al.*, 2006a) allowing for greater environmental interaction than previously, indicating a greater potential for warming before reaching the gauge. However through the comparison of Gornera and Massa, temperature changes may be increased but warming rates over distance may remain similar to those found in this study, as discharge levels will be elevated and with more flow to heat changes in short-wave radiation and air temperature may not be substantial enough to offset the control on water temperatures.

Tributary glaciers make up large areas of valley glaciers and under a period of persistent recession will become detached from the main trunk glacier and stand-alone creating discharges which flow out from the snout before joining bulk meltwaters from the primary glacier. These meltwaters will initially be heated in the environment before re-entering the subglacial system of the trunk glacier and once again emerging from the portal (Jansson *et al.*, 2003); it is unclear as to whether these waters will be returned to equilibrium levels close to 0°C or is capable of increasing the emerging initial water temperature.

Tributary glaciers currently contribute to the discharge of the Massa; stream flow travels 2km before entering main channel. Detached tributary glaciers in the Gornergletscher system are much smaller than those within the Aletschgletscher system; therefore produce much less of total annual runoff. Generated runoff travels at a slower velocity allowing for greater interaction with incoming solar radiation resulting in larger increases in stream water temperature; however the lack of volume of meltwater is likely to offset any potential increase. Uehlinger *et al.* (2003) demonstrated variations in water temperatures across a glacial floodplain where discharges were lower than the main channel flow, finding greater increases in slower moving small streams. Meltwater from the tributaries of Gornergletscher re-enter the subglacial system as the glacier extends beyond currently detached tributaries meaning that flow from these glaciers is likely to be altered from its input temperature to one closer to the mean of bulk flow (0.1°C), therefore the effects of prior heating are likely to be only minor meltwater generation as the warmer flow contacted with the ice mass (Rothlisberger and Lang, 1987).

5.4 Model Design

The concept of modelling water temperature in bodies of water is one dating back to the 1970s (Caissie, 2006). As water temperature has such a profound impact upon the habitats of aquatic organisms there is a necessity to be able to predict any future changes to habitats and counter negative effects, where possible. Modelling of temperature change in glacially-sourced streams is complex; primarily due to enhanced levels of meltwater generation with increases in air temperature and solar radiation, which reduces the potential for temperature change. Models relating to the proglacial system must therefore account for changes which occur within the glacial system at different air temperatures. Therefore, it is proposed that alongside a water temperature model, a degree day (temperature-index) model should be used to simulate melt generation and subsequent discharge levels.

5.4.1 Stream water temperature modelling

Different approaches of varying sophistication have been developed to model aspects of thermal stream behaviour in space and in time (Moore, 2005; Caissie, 2006). Models have been designed at a temporal scale to cover timescales from hourly to annually; modelling at differing timescales highlights changes in the relationship between energy availability and stream water temperature. Spatial variations in water temperature models range from short specific segments of a river (e.g. SSTEMP) up to complete watershed and stream network coverage (e.g. SNTEMP) (Bartholow, 2002). These approaches allow modellers to understand stream water temperatures in catchments at micro and macro scales.

5.4.2 Stochastic & Regression Modelling

Stochastic and regression models are computationally simple and as such are much easier to compile and understand than models that are more complex. Stochastic and regression models have included empirical models that rely on statistical analysis to make predictions from weather data or information on catchment characteristics (Webb *et al.*, 2008). They are also applicable where air temperature data are available at locations. However, these approaches are no guarantee that seasonality of data is completely removed to achieve stationary residuals (Benyaha *et al.,* 2008), because the relationship developed between variables may only be related to the measured period. Using methods such as harmonic analysis (e.g. Caissie *et al*., 1998; 2001) and the use of equilibrium temperature concepts (e.g. Bogan *et al*., 2004; Caissie *et al*., 2005), deviations from seasonal trends can be identified and accounted for (Webb *et al.,* 2008). Examples of empirical models, which rely on statistical (regression) analysis to make predictions from weather data or information on catchment characteristics, are: Stoneman and Jones, 1996; Donato, 2003; Neumann *et al*., 2003; Rivers-Moore and Lorentz, 2004; Benyahya *et al*., 2007.

The equilibrium temperature method attempts to better understand the thermal dynamics of the river system by interpreting the relationship between the total heat fluxes at the water surface that is proportional to the difference between the water temperature and an equilibrium temperature (Te) (Caissie, 2005). It has been shown that the equilibrium temperature crosses hourly water temperatures twice daily and over an annual period twice, once on the warming trend and then again on the cooling trend (Edinger *et al*., 1968).

Gu *et al*. (1998) used the equilibrium temperature method to quantify the effect of river discharge variations on river thermal regime. Good agreement was illustrated between time-average stream water temperature data as a function of discharge. Diurnal amplitudes of water temperature were related to diurnal amplitudes of equilibrium temperature, which followed an exponential decay with increasing discharge. They concluded that by using the equilibrium temperature, increased river discharge through water releases below dams can present a good opportunity to reduce river water temperature.

5.4.3 Deterministic Modelling

Approaches to modelling ranging from a simple regression model to a deterministic or energy budget model that can be applied to predict stream water temperatures. A deterministic modelling approach can be quite elaborate because it can be capable of calculating all relevant heat fluxes at both the water surface and sediment water interface (Caissie *et al.,* 2005). Results from the sediment water interface are only of consequence in shallow streams. Deterministic models are demanding in terms of model development and data requirement, which makes a simpler model more attractive. Deterministic models are well adapted to study the impacts of flow reduction and/or flow alteration on the river thermal regime (Caissie *et al.*, 2007), and will prove useful in climatic change scenarios, which dictate a change in runoff patterns in high mountain basins.

5.4.4 Degree-day Modelling

Melt modelling of snow and ice is an essential tool in attempting to predict runoff from nival and glacierised basins. Long has it been recognized that air temperature changes are only relevant to glacier melting at temperatures close to, or above, the melting point of ice (0°C) (Braithwaite *et al.*, 2013). Degree-day models or temperature-index models have been a common method for modelling melt primarily due to wide availability of air temperature data and, consequent general good model performance (Correlation coefficient =0.96, Braithwaite and Olesen (1989)) and simple interpolation (Hock, 2003). Temperature-index models generally match energy balance models in terms of performance at catchment scale (WMO, 1986). Air temperature correlates highly with many energy balance components and as such offers a simpler accurate alternative to using net radiation (Hock, 2003).

Basin hypsometry (distribution of basin area with elevation) is an important external factor in determining a glaciers response to climatic change (Small, 1995). Hypsometry interacts with Equilibrium Line Altitude (ELA), transient snowline and 0°C isotherm determining a glaciers accumulation/ ablation area ratio (Collins, 1998). Basin sensitivity is dependent on climatic patterns; basins situated in maritime conditions (New Zealand) are highly sensitive to changes in air temperatures and precipitation. In comparison, basins in Polar Regions receive little input from precipitation but experience little air temperature fluctuation (Figure 5.7).

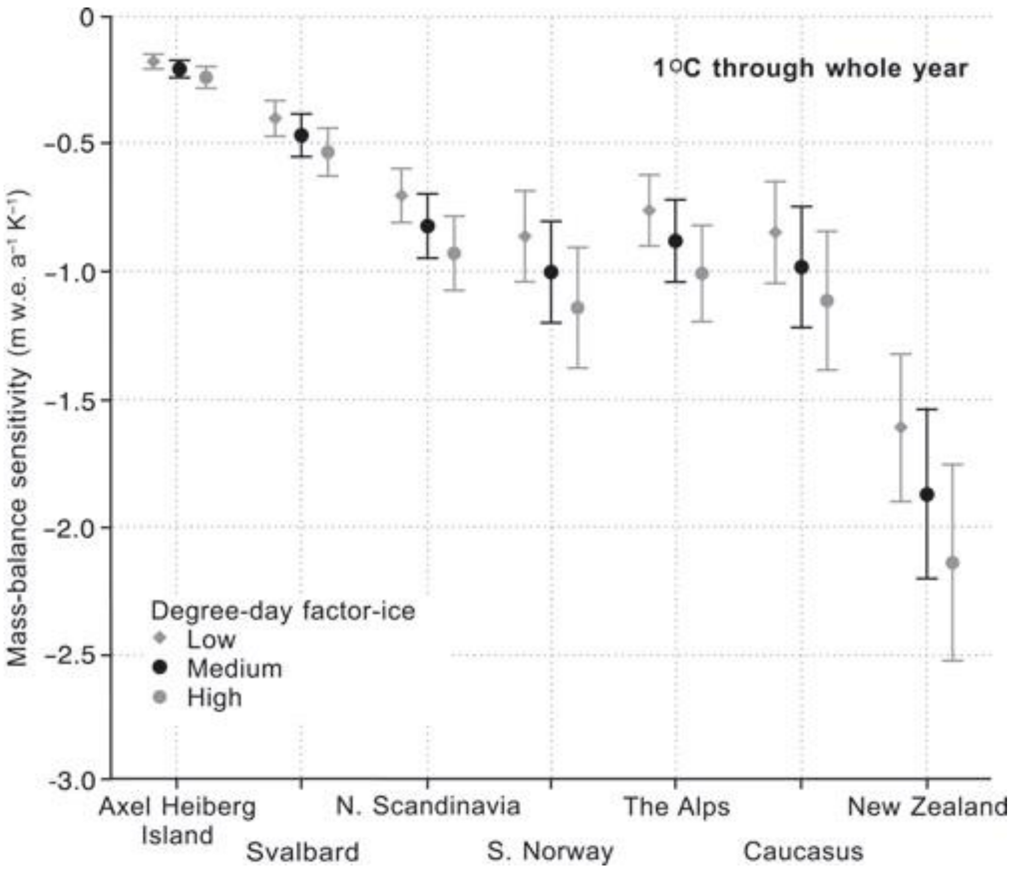

Figure 5.7- Mass-balance sensitivities for seven glacial regions computed with a degree-day model applied to estimated balanced-budget ELAs within each region. Low, medium and high degree-day factors are 6, 7 and 8mmd^{-1}K^{-1} (Braithwaite *et al.*, 2013)

5.4.5 Model Selection

The author wishes to make recommendations which will assist in predicting future patterns of proglacial stream water temperature in response to a warming climate. In order to achieve this, two aspects require modelling, the amount of runoff generated from the glacier in a warming climate and potential water temperature changes experienced by the discharge entering the proglacial system between the portal and gauging station. A glacial temperature index runoff model can be implemented in order to predict future levels of mean daily snow and ice derived runoff in warmer climatic conditions and a water temperature model, ought to be used to predict mean daily temperature changes between glacier portal and gauging station in modelled future runoff. The Stream Segment Temperature Model (SSTEMP) is a model which is capable of simulating proglacial stream temperatures, as it operates under Lagrangian theory tracking an aliquot of water as it moves through time and space.

Stream Segment Temperature Model (SSTEMP) is an adaptation of the Stream Network Temperature Model (SNTEMP) which follows an aliquot (parcel) of water as it moves through a predefined length of stream. SSTEMP differs from SNTEMP, which models large networks of rivers and their tributaries, as it can map the effects of reservoir release (Bartholow, 2002). Glaciers are essentially reservoirs storing millions of litres of water, and under certain climatic conditions enter steady state release of flow at a uniform temperature akin to conventional man-made reservoirs or dams. Both SSTEMP and SNTEMP are examples of a deterministic approach to water temperature modelling.

SSTEMP predicts mean, maximum and minimum daily water temperatures at specified distances downstream, using inputted data which describe meteorology, hydrology, stream geometry and optionally, stream shading (Bartholow, 2002). These data can be simply measured at study sites. SSTEMP simulates the various heat flux processes, which determine positive and negative temperature changes as an aliquot of water moves through the stream segment. Convection, conduction, evaporation and long- and short-wave radiation all affect the magnitude and rate of temperature increases in a stream. SSTEMP determines the amount of short-wave radiation intercepted by topographic and riparian shading. The water temperature model contains a reservoir function, which allows for atmospheric interactions with the flow to begin when an aliquot first enters the system. Without this function waters entering a system would be presumed to be reacting to upstream atmospheric parameters based on those set for the stream segment (Bartholow, 2002).

A simple temperature index model which calculates runoff levels for days where mean daily air temperatures are above 0°C, can be used to simulate future levels of runoff in glaciated basins. To account for spatial variability in melt rates, while using spatially constant degree-day factors, models divide the basin into elevation bands to consider a decrease in melt levels with increasing elevation (Hock, 2005). Elevation bands indicate amounts of total ice cover at defined elevations, and melt levels are adapted using the adiabatic lapse rate for determining temperature change at higher elevations. Degree-day factors (DDF) are calculated for individual basins either by using direct measurements from ablation stakes or from melt obtained by energy balance calculations (Hock, 2003). The suggested DDF for Gornergletscher is 6mm d^{-1} $°C^{-1}$ (millimetres of melt per day) (Collins, personal communication) and 11.7mm d^{-1} $°C^{-1}$ for Aletschgletscher (Lang, 1986). To determine model accuracy air temperature from previous years can be used to simulate discharge levels which can then be correlated with actual corresponding discharge to determine accuracy.

SSTEMP is used primarily to accurately model water temperatures at a daily scale. However, it can be used to model hourly and monthly water temperatures (Bartholow, 2002). Used alongside a temperature index model daily mean temperatures and flows ought to be used as they are the most accurate. This method ought not to be used in streams where data indicate an unpredictability of flow patterns over short timescales; it is therefore not wise to use the SSTEMP model for modelling the impacts of a rapid drainage event from the Gornersee at Gornergletscher.

Temperature-index models can be easily adapted to simulate future warmer and cooler periods, and as such can prove to be ideal for simulating glacier runoff responses. SSTEMP is robust enough to deal with high discharges as long as site specific data parameters are inputted accurately. SSTEMP requires good quality data in order to accurately ascertain accurate model results. Alpine Glacier Project and BAFU has collected hourly data of discharge and in-stream and climatic variables of Gornera and Massa basins in the Swiss Alps since 1970. This data provides a comprehensive representation of multiple different climatic scenarios over a 40-year period, which can be used to accurately predict and project future scenarios within an ever-changing climate.

6. Conclusions

6.1 Summary of key findings

Highly glaciated basins have thus far been avoided in terms of water temperature studies with the focus being placed upon small scale glaciated basins (Brown *et al.* 2006a, Brown *et al.* 2006b, Cadbury *et al*. 2008), where discharges are low and environmental changes have a greater effect.

This study has documented evidence of a spring maximum temperature much greater than the annual average, previously unseen in other water temperature studies. The extent of the spring warm water pulse is reliant on high short-wave radiation conditions and the amount of snow from previous accumulation cycles. Where snow accumulation levels are above average, high albedo snowpack remains on the glacier for a longer period of time delaying the bulk of runoff entering the channel, allowing water temperatures to reach their annual peak for longer periods. High short-wave radiation which coincides with below average snow accumulation leads to a shorter period of high temperatures as ice melt begins much earlier.

Diurnal regimes of water temperature across both basins were similar with environmentally derived change replicated in Massa and Gornera. The level of temperature change is much greater in Massa as the stream segment between portal and gauge is much larger; however observed warming rates over the ablation season are lower than Gornera and both basins warm less than previous studies which have focussed on smaller glacierised areas. This indicates that glacier size directly influences the level of heating of water courses issuing from the portal, with large glaciers having the greatest reducing the level water temperature change.

The level of water temperature change is primarily controlled by discharge during the peak of the ablation season; as high incoming short-wave radiation levels and air temperatures create large cold water pulses which offset the ability for greater surface-atmosphere interactions. Diurnal data show that water temperature is prevented from rising further during afternoons due to the delayed cold-water pulse of discharge generated from earlier melting. The lag exists due to meltwater being sourced at the head of the catchment.

As discharges from highly glacierised basins are predicted to initially rise, with a warming climate, before falling away in the long term as the percentage of annual flow from icemelt decreases and large glaciers occupy less area and percentage of the basin. Annual discharge patterns will then be dominated by seasonal snowmelt, summer precipitation and groundwater flow. In the period where discharges from icemelt are forecasted to increase, water temperatures are likely to remain stable and fluctuate at observed levels during the ablation season. Yet outside this period early season maximums of water temperature ought to be expected to increase, as the lowest elevations of the glacier will be further from the initial transient snowline, therefore initial icemelt will be delayed. This can be countered by the argument that as short-wave radiation levels and air temperatures will be higher the period will be shorter as snow is melted at a much greater rate, reducing the warm water pulse and increasing the length of the ablation season.

6.2 Further Research

An annual hourly water temperature measurement system should be created at the gauging station on the Gornera in order to establish whether annual discharge patterns seen in the Massa are a local phenomenon or more wide spread through other highly glacierised Alpine basins. This system will also nullify gaps in the data caused by the emergence of temporary probes from low discharges, especially given that low discharges show the greatest variation in water temperature.

As continued glacier recession is predicted, the inflow of waters from tributary glaciers, detached from the main trunk glacier, should be monitored to understand the effects of the difference in water temperatures and the effects on main flow. Tributary glaciers will produce lower levels of discharge than the main ice body; therefore velocity will be reduced increasing the residence time and interactions with the streambed and environment allowing for greater temperature changes.

Due to the complexities of the relationship between discharge generation through ice and snowmelt and the subsequent effects upon water temperature, water temperature models are unable to accurately assess the relationship of high discharge and high incoming short-wave radiation/ air temperatures. In order to rectify this issue, the author makes suggestions (5.4) for using a positive degree day model alongside a stream segment water temperature model to predict mean daily temperature. This method would allow for future climatic situations to be modelled where air temperatures are higher and the area of ice decreased, however the method can only be used where waters flow in a single channel.

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